

Improved Tropical Cyclone Rapid Intensification Forecasts in Current and Future Forecasting Systems

DR. PETER M. FINOCCHIO

*Atmospheric Dynamics and Prediction Branch
Marine Meteorology Division*

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EXECUTIVE SUMMARY

This report presents research conducted by Peter M. Finocchio as part of a Karles Fellowship at the Naval Research Laboratory. This research covered two topics that are highly relevant to battlespace environments: the rapid intensification of tropical cyclones and the effect of Arctic summer cyclones on sea ice. The first part of this research on the rapid intensification of tropical cyclones involved implementing a series of idealized numerical simulations aimed at understanding how increasing vertical wind shear in the environment around a tropical cyclone affects rapid intensification differently depending on the stage of rapid intensification when the shear increases. The second part of this research is a statistical analysis of a large set of Arctic summer cyclones and the subsequent changes in sea ice after cyclone passage. Both parts of this project resulted in three peer reviewed publications (Finocchio and Rios-Berrios 2021, Finocchio and Doyle 2021, Fearon et al. 2021). These publications provide critical new understanding of cyclones and their impacts on Navy operations that will guide the development of Naval prediction systems and facilitate forecasting efforts in both the tropics and the Arctic.

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IMPROVED TROPICAL CYCLONE RAPID INTENSIFICATION FORECASTS IN CURRENT AND FUTURE FORECASTING SYSTEMS

1. INTRODUCTION

1.1 The Effect of Increasing Vertical Wind Shear on Tropical Cyclone Rapid Intensification

1.1.1 Naval Relevance

This research addresses the primary objective of this Karles Fellowship, which is to improve our understanding and ability to forecast the rapid intensification (RI) of Tropical Cyclones (TCs). RI is not well predicted by Naval environmental prediction systems such as COAMPS-TC, which is part of the reason that TCs remain a considerable threat to U.S. Navy operations both near shore and at sea. Recent examples of RI occurring just before a TC makes landfall in the United States – such as Hurricane in Michael in 2018, which caused more than [\\$648 million worth of damage of Tyndall Air Force Base](#) – highlight the urgency for improved TC intensity forecasts. As such, improving predictions of TC intensification – especially RI – is a top priority of mesoscale modeling efforts at the Naval Research Laboratory. This research specifically addresses the role that vertical wind shear in the environment around a TC affects its intensification, with a focus on RI. The findings of this work have direct applicability to future Navy modeling efforts to improve TC intensity change forecasts and to better harness operational Navy prediction systems like COAMPS-TC and the COAMPS-TC ensemble in order to provide improved TC guidance to the fleet.

1.1.2 Scientific Background

The vertical wind shear in the environment around a tropical cyclone (TC) is a leading cause of TC intensity change. In most cases, stronger vertical wind shear is detrimental to intensification (DeMaria and Kaplan 1994), but some TCs are able to rapidly intensify in environments with moderate-to-strong vertical wind shear magnitudes (e.g., Rios-Berrios et al. 2016a,b; Ryglicki et al. 2018). The intensity and structure of a TC can determine the extent to which a TC resists the destructive effects of vertical wind shear, but the reasons for this are not fully understood and have not been explored in detail. Therefore, the purpose of this study is to systematically examine how TC intensity and size modulate the storm's intensity response to increasing vertical wind shear using idealized, full-physics numerical simulations.

A key part of the TC response to vertical wind shear is the development and evolution of vortex tilt. Jones (1995) showed that sheared barotropic vortices initially tilt downshear, and then precess cyclonically toward a quasi-steady left-of-shear tilt orientation, in which the cyclonic precession counteracts differential advection by the sheared background flow. A vertically aligned vortex is more likely to intensify than a tilted vortex because the positive feedbacks between near-surface convergence and latent heat release above the surface circulation that spin up the surface winds are most efficient when a vortex is vertically aligned (Nguyen et al. 2008, Rios-Berrios et al. 2018). Therefore, understanding the processes involved in either the re-establishment or maintenance of vortex alignment is central to understanding intensity changes of sheared TCs.

Several hypotheses for vortex re-alignment have been proposed. One hypothesis is that re-alignment can occur through dry, inviscid damping of a wavenumber-1 quasi-mode excited by the vertical wind shear (Reasor and Montgomery 2001, Reasor et al 2004, Reasor and Montgomery 2015). Damping of this mode may be enhanced in the presence of cloud processes (Schecter and Montgomery 2007). Recent modeling studies utilizing fully moist simulations demonstrate a pathway to vortex re-alignment that essentially extends the ideas of Jones (1995), whereby a tilted vortex that precesses upshear realigns through differential advection by the sheared background flow (Tao and Zhang 2014, Tao and Zhang 2015, Finocchio et al. 2016). Deep convection downshear and humidification upshear are thought to promote vortex alignment by enhancing upshear precession and the symmetrization of precipitation (Tao and Zhang 2015, Zawislak et al. 2016, Rogers et al. 2016, Rios-Berrios and Torn 2017, Alvey et al. 2020). Rios-Berrios et al. (2018) and Rios-Berrios (2020) proposed a new vortex realignment pathway for weak TCs that acknowledges the importance of moist diabatic processes and does not necessarily involve the upshear precession of a coherent vortex. Through this pathway, a vertically aligned vortex can emerge from axisymmetrization of convectively generated vorticity within a mid-tropospheric meso-vortex that favors recirculation of moist air parcels.

Vortex re-alignment and TC intensification are disrupted when storm-relative flows caused by vertical wind shear bring relatively dry and cool air into the TC inner core (i.e., ventilation). Ventilation is hypothesized to weaken TCs by diluting the upper-level warm core (Gray 1968, Frank and Ritchie 1999, Kwon and Frank 2008, Fu et al 2019) and lowering the entropy within the eyewall via horizontal eddy fluxes (Simpson and Riehl 1958, Cram et al. 2007, Tang and Emanuel 2010, 2012, Alland et al. 2021b). Shear-induced downdrafts also flush the boundary layer with low-entropy air, which can be entrained into eyewall updrafts via the low-level inflow (Riemer et al. 2010, 2013, Tang and Emanuel 2010, 2012, Alland et al. 2021a).

Although the extent to which vertical wind shear tilts and ventilates the TC vortex is known to depend on environmental humidity and the wind shear magnitude (Tang and Emanuel 2010, Riemer and Montgomery 2011, Braun et al 2012, Ge et al. 2013, Alland et al. 2021a,b), it may also depend on characteristics of the TC (DeMaria et al. 2005). Jones (1995) found that broader and more intense vortices tilt less in response to shear than weaker and smaller vortices due to stronger vertical coupling that promotes cyclonic precession. Reasor et al. (2004) and Reasor and Montgomery (2015) found that a vortex with a faster radial decay of the tangential winds outside of the inner core evolves more rapidly toward a stable left-of-shear tilt orientation, which was confirmed in full-physics simulations of mature TCs (Riemer and Montgomery 2013). Furthermore, Riemer and Montgomery (2011) showed in a kinematic analysis that, for a given storm-relative horizontal flow, the rate at which environmental air flows into the TC inner core decreases as the strength of the tangential winds increases. Stronger TCs may also be more resilient to ventilation by shear-induced downdrafts. With all else being equal, the stronger surface winds of a more intense TC increase the enthalpy flux from the sea surface, enabling a faster recovery of downdraft air parcels within the boundary layer (Powell 1990, Zhang et al. 2013, Nguyen et al. 2019).

While several studies have touched upon the notion that the TC response to vertical wind shear depends to some extent on the size and/or intensity of the storm, it is the primary purpose in this study to systematically investigate this dependence. To this end, we use idealized simulations in which the vertical wind shear gradually increases from weak to strong magnitudes at different points in the development of a TC in an otherwise favorable environment.

1.2 Cyclone Impacts on Sea Ice in the Arctic during Summer

1.2.1 Naval Relevance

Although the primary focus of this Karles Fellowship was on Tropical Cyclone Rapid Intensification, a significant portion of the work performed involved participation in the ONR Arctic Cyclone Departmental Research Initiative. Substantial contributions to this program were made during

the fellowship on understanding how cyclones in the Arctic affect sea ice cover on the 1-10 day timescales that are critical for Naval deployment and navigability within the Arctic. All of the material in this report about this work appear in a recent article published in a special issue of *Frontiers in Earth Science* (Finocchio and Doyle 2021).

1.2.2 Scientific Background

Arctic summer sea ice cover is decreasing at such a rate that ice-free summers are projected to occur by the middle of this century (Stroeve et al. 2012, Overland and Wang 2013). This remarkable loss of sea ice is a result of the significant warming of the Arctic atmosphere and ocean that has taken place in recent decades. Whether transient Arctic cyclones during the summer months have contributed to the accelerating loss of sea ice remains unclear. There has been no significant increase in Arctic-wide cyclone frequency or intensity that would suggest stronger or more frequent cyclones during the summer are responsible for accelerating summer ice loss (Screen et al. 2011, Koyama et al. 2017, Semenov et al. 2019). However, cyclones are likely having a larger effect on sea ice cover as young and thin sea ice becomes more widespread in the Arctic (Rampal et al. 2009, Maslanik et al. 2011, Spreen et al. 2011, Lindsay and Schweiger 2015, Itkin et al. 2017). Despite uncertainty in whether cyclones are a significant source of seasonal or interannual variability in pan-Arctic sea ice cover (Rae et al. 2017), numerous studies have demonstrated that cyclones are indeed a source of local and short term variability in sea ice cover and concentration (Zhang et al. 2013, Stern et al. 2020, Lukovich et al. 2021), and these short-term changes in sea ice are critical for navigability within the Arctic region. This study examines how the surface winds and atmospheric energy flux from recent Arctic summer cyclones affects sea ice concentration on synoptic weather time scales of 1-10 days.

The strong surface winds associated with summer cyclones affect sea ice in several ways. Wind-induced drift is a key source of short term sea ice variability – particularly where the sea ice cover is thinner, farther from the coastline, and primarily composed of floes (Thorndike and Colony 1982, Hakkinen et al. 2008, Spreen et al. 2011, Vihma et al. 2012). Whether the wind-induced advection of sea ice increases or decreases sea ice concentration at a given location, however, depends on the direction of the surface winds relative to the local gradient in ice concentration. Cyclonic surface winds enhance sea ice divergence, which locally reduces sea ice concentration in the vicinity of a cyclone (Maslanik and Barry 1989, Kriegsmann and Brummer 2014, Lukovich et al. 2021). In addition, strong surface winds later in the summer are increasingly capable of mixing relatively warm sub-surface sea water upward toward the surface, which causes substantial melting along the ocean-facing side of sea ice (Steele et al. 2010, Zhang et al. 2013, Stern et al. 2020).

Cyclones also contribute to the short-term variability in sea ice concentration by changing the local energy balance on the atmosphere-facing side of sea ice. Clouds associated with cyclones reflect solar radiation that would otherwise melt the top surface of sea ice and/or heat the open water areas around sea ice (Finocchio et al. 2020, Schreiber and Serreze 2020). Finocchio et al. (2020) found this cloud shading effect was particularly strong in the Arctic marginal ice zone (MIZ) in June, when shortwave radiation reaches its annual maximum and when clouds have a strong cooling effect at the surface (Kay and L'Ecuyer 2013). Schreiber and Serreze (2020) showed that the cloud shading effect of cyclones temporarily decelerates the seasonal reduction in sea ice concentration near the ice edge during summer. Although cloud cover also increases the downward longwave flux at the surface, which can have important implications for sea ice change in the winter and spring months (Shupe and Intrieri 2004, Persson 2012), Finocchio et al. (2020) found that the average reduction in shortwave radiation due to early summer cyclones more than compensates for the increased longwave radiation from clouds in their sample of cyclone cases, resulting in a net decrease in the atmospheric energy flux at the surface due to cyclones. Although Arctic cyclones tend to decelerate the seasonal loss of sea ice early in the melt season, the strong sensible and latent heat fluxes within the warm sector of cyclones later in the melt season can locally accelerate sea ice loss. For example, Stern et al. (2020) showed that the sensible heat flux in the warm sector of an exceptionally strong cyclone in August 2012 (frequently referred to as the ‘Great

Arctic Cyclone') contributed to anomalous surface energy forcing that locally accelerated ice loss in the Beaufort Sea.

Cyclone impacts on sea ice vary from case to case not only because of variability in the cyclones themselves, but also because of differences in the underlying sea ice cover. Lukovich et al. (2021) compared the impacts of similarly intense Arctic cyclones in August 2012 and August 2016 on the sea ice. They found that the August 2012 cyclone had a more destructive effect on sea ice primarily because it affected thinner, lower-concentration sea ice than the August 2016 cyclone. The apparent dependency of cyclone impacts on the characteristics of the underlying sea ice motivates a new approach to analyzing the relationship between cyclones and sea ice loss that involves stratifying cyclone impacts based on the underlying sea ice concentration.

Similar to Finocchio et al. (2020), this study adopts a statistical approach to analyzing the effects of Arctic cyclones on sea ice within the marginal ice zone. We focus here on the Pacific sector of the Arctic, which experiences high interannual variability in summer sea ice cover and frequent summer cyclones. Within this region, we stratify the effects of cyclones on sea ice by the sea ice concentration in order to gain a more detailed understanding of the relationship between cyclone impacts and short-term sea ice changes.

2. APPROACH

2.1 The Effect of Increasing Vertical Wind Shear on Tropical Cyclone Rapid Intensification

The simulations for this part of the research conducted as part of this Karles Fellowship utilize version 3.9.1.1 of the compressible, nonhydrostatic Weather Research and Forecasting model with the Advanced Research dynamical core (WRF-ARW; Skamarock et al., 2008). The model is configured with 60 equally spaced mass levels extending from the sea surface to 20 km altitude. The sea surface temperature is held constant at 28°C throughout the simulations. We use three domains: a stationary domain and two nested domains that follow the vortex minimum geopotential height at 700 hPa. The outer domain is a doubly periodic f -plane at 20°N with 18-km grid spacing and square dimensions of 4,320 km. The nested grids employ two-way nesting and have 6-km and 2-km grid spacing with square dimensions of 1,440 km and 720 km, respectively. We use the Thompson microphysics scheme, which is double moment in cloud ice and single moment in rain, snow, graupel, and cloud water (Thompson et al. 2008). We also use the YSU planetary boundary layer scheme (Hong et al. 2006). For simplicity, we do not use cumulus or radiation schemes. Horizontal diffusion is computed along model levels from the horizontal deformation ($\text{diff_opt}=1$ and $\text{km_opt}=4$ in the namelist).

We first conduct a 240-h baseline simulation in which a weak initial TC evolves into a mature hurricane in a favorable environment with weak vertical wind shear. The initial TC is a cloud-free, Rankine vortex with a maximum tangential wind speed of 15 m s^{-1} that is located at a radius of 90 km and a height of 1.5 km. The winds decay exponentially away from the radius of maximum winds (RMW) with a radial decay exponent $\alpha=0.4$. The mass fields that are in gradient wind balance with the prescribed wind field are obtained through an iterative procedure described in Nolan et al. (2001). The environment around the vortex has the moist tropical temperature and humidity sounding from Dunion (2011), and a zonal wind profile with 3 m s^{-1} of westerly vertical wind shear between 200 and 850 hPa (Fig. 1a, dashed line). The meridional wind speed of the sheared flow is initially 0 m s^{-1} at all levels. Following Nolan (2011), we use the “perturbation Coriolis” technique in WRF, in which the Coriolis torque only acts on perturbations to the specified wind profiles. This allows us to initialize vertical wind shear in a doubly periodic domain without meridional temperature gradients. In addition, the temperature (T), moisture (q), and winds (u , v) are nudged to specified profiles (Stauffer and Seaman 1990). Nudging adds a Newtonian relaxation term to the WRF tendency equations for each state variable:

$$\frac{\partial V(x,y,z,t)}{\partial t} = \dots - \frac{V(x,y,z,t) - V_{ref}(z,t)}{\tau} \quad (1)$$

in which V is the state variable being nudged, V_{ref} is the reference profile toward which the state variable is nudged, and τ is the relaxation time scale (larger τ results in weaker nudging). For the baseline simulation, we set $\tau = 36$ h for all four of the nudged state variables and nudge to the profiles specified above throughout the entire simulation. Nudging is only applied in the free troposphere (above the boundary layer) and in the 18-km domain. Nudging the large-scale environment to the specified wind profile can hinder a TC's natural ability to weaken the shear flow in the vicinity of the TC, but we believe that our use of large vortex-following domains in which nudging is not applied alleviates this issue.

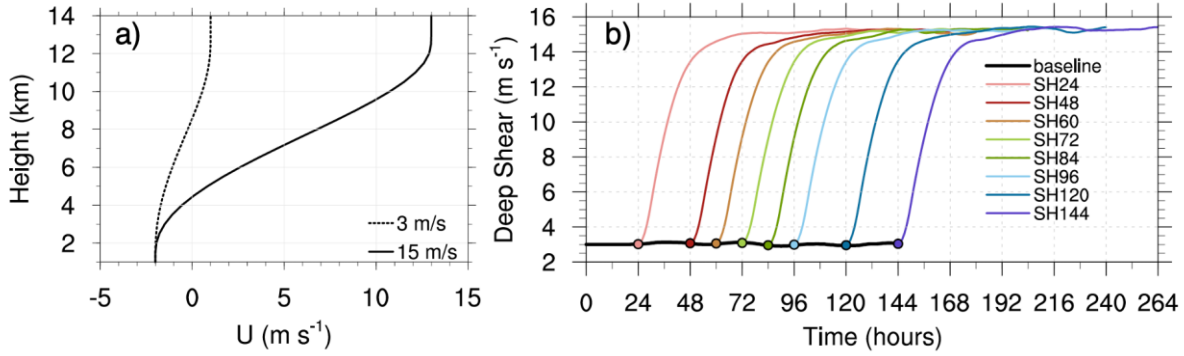


Fig. 1 — (a) Profiles of zonal wind speed imposed in the outer domain. The 3-m s⁻¹ baseline shear profile is the dotted line and the target 15-m s⁻¹ shear profile is the solid line; (b) time series of deep-layer (850–200-hPa) vertical wind shear averaged along the edge of the outer domain of each simulation.

The thick black line in Figs. 2a,c shows the evolution of minimum sea-level pressure (Fig. 2a) and maximum azimuthally averaged tangential wind speed at 10-m (Fig. 2c) for the TC in the baseline simulation. The TC starts to slowly intensify after 24 h, and then undergoes a period of rapid intensification (RI) starting at 46 h and lasting until 100 h (RI is defined as an increase in the maximum wind speed ≥ 30 kt in 24 h). The maximum wind speed increases from 28 to 65 m s⁻¹ (54–126 kt) during RI, and continues to steadily increase after RI to a maximum value of 71 m s⁻¹ (138 kt) at 125 h. From 125–144 h, the maximum wind speed weakens and the 34-kt (50-kt) wind radius expands by 30 km (18 km).

We then conduct eight 120-h simulations that increase the westerly vertical wind shear in the outer domain from 3 m s⁻¹ to 15 m s⁻¹ at different points in the baseline simulation. The target zonal wind profile with 15 m s⁻¹ shear (Fig. 1a, solid line) is similar to the profile with 3 m s⁻¹ shear except for having 13-m s⁻¹ westerlies above 200 hPa. The environmental surface winds are left the same to ensure that any change in surface fluxes is due only to the changing TC. We use the time-varying point downscaling technique, which nudges the outer domain to a pre-defined sequence of large-scale environments that change over time in order to smoothly increase vertical wind shear around the TC during the simulations (Onderlinde and Nolan 2017). This technique has been used in recent TC modeling studies (e.g., Finocchio and Majumdar 2017b, Alland et al. 2021a,b) to avoid having to spin up an initially cloud-free vortex in strong shear or impose a step-function increase in shear on a more developed TC. The relaxation time scale (τ in Eq. 1) controls the rate at which state variables change during the simulations. In order to increase shear from 3 to 15 m s⁻¹ fast enough to allow the TC to respond to the stronger shear within the 120-h simulations, we had to decrease τ for the u and v winds from 36 to 12 h upon restarting the model while keeping $\tau = 36$ h for T and q . With this setup, the shear in the outer domain reaches the target value of 15 m s⁻¹ after about 36 h (Fig. 1b). We found in several test simulations that this setup results in the smoothest transition of the wind and thermodynamic fields in all three model domains after restarting the model from the baseline simulation. Using a nudging time scale in the increasing shear simulations that

differs from the baseline simulation, however, makes it difficult to attribute differences between the baseline and increasing shear simulations to differences in the vertical wind shear. Therefore, we implement a parallel simulation for each increasing shear simulation that maintains the 3-m s^{-1} shear from the baseline simulation, but uses the shorter 12-h nudging time scale on the horizontal winds from the increasing shear simulations.

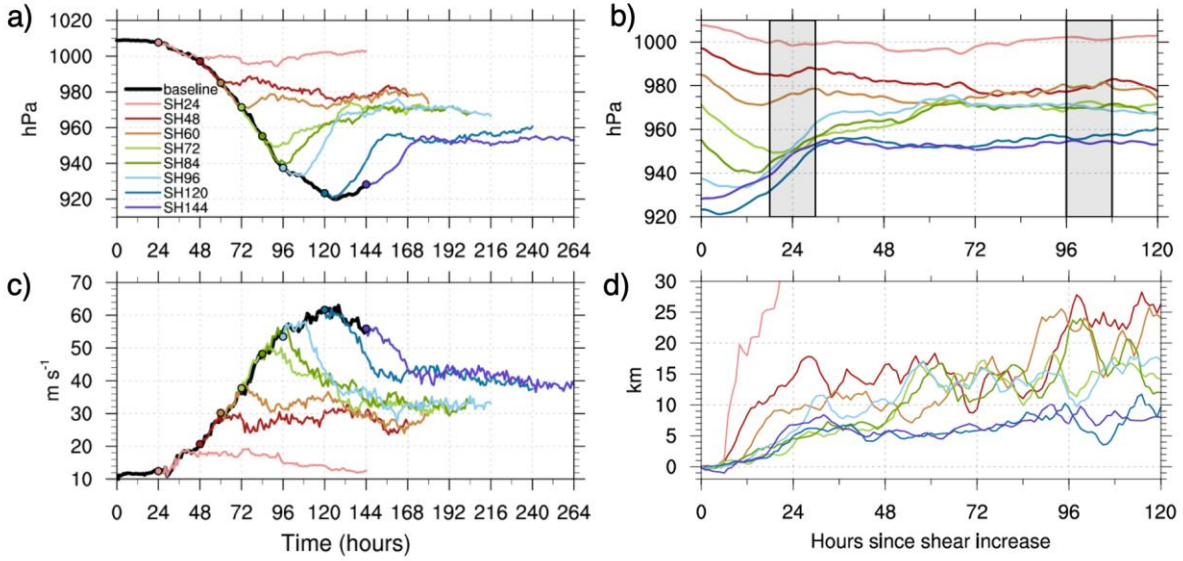


Fig. 2 — Times series of (a) TC minimum sea-level pressure and (c) maximum azimuthally averaged 10-m wind speed; (b) as in (a), but only for the increasing shear simulations and plotted relative to the time when shear starts increasing; (d) time series of the differences in 1–6 km vortex tilt between the increasing shear simulations and the parallel simulations with constant 3-m s^{-1} shear. The gray shaded regions in (b) denote the initial weakening and near-steady intensity periods that are the focus of the analysis.

The eight increasing shear simulations are listed in Table 1 along with the TC intensity averaged during the first 6 hours after shear starts increasing. We refer to the increasing shear simulations by “SH” followed by the time at which the shear starts to increase (e.g., “SH24” is the simulation in which vertical wind shear starts to increase 24 h into the baseline simulation). The top row of Fig. 3 shows the simulated radar reflectivity in five of the increasing shear simulations averaged during the first 6 hours after shear starts increasing. In SH24, shear increases when the TC is a tropical storm with convection localized downshear. The TC in SH48 is also a tropical storm, but has just begun RI and has a more symmetric precipitation structure than in SH24. In SH60–SH84, the shear increases as the TC rapidly intensifies from a Category 1 (SH60) to a Category 4 hurricane (SH84) and develops an annular precipitation structure. In SH96–SH144, the shear increases while the TC is a Category 4 hurricane with an annular ring of intense convection.

Table 1 — Maximum 10-m wind speed and minimum sea-level pressure (MSLP) averaged during the first 6 h after shear starts increasing and during the last 12 h of each simulation. The category from the Saffir-Simpson scale corresponding to each maximum 10-m wind speed appears in parentheses. (Right two columns) Vortex tilt from 1–6 km averaged from 18–24 h after shear starts increasing and during the last 12 h of the each simulation.

	Max 10-m wind speed (m s^{-1})		MSLP (hPa)		1–6 km Vortex tilt (km)	
	First 6 h	Last 12 h	First 6 h	Last 12 h	18–24 h	Last 12 h
SH24	24.2 (TS)	27.4 (TS)	1006.8	1002.3	39.7	336.2
SH48	28.6 (TS)	40.5 (C1)	994.5	981.0	15.0	26.0
SH60	38.5 (C1)	42.5 (C2)	981.6	977.8	10.1	24.0
SH72	49.4 (C3)	44.6 (C2)	966.2	970.8	3.2	16.8
SH84	60.4 (C4)	42.2 (C2)	949.8	969.6	3.9	17.8
SH96	65.0 (C4)	43.8 (C2)	935.5	967.8	4.8	18.2
SH120	69.1 (C4)	52.1 (C3)	922.3	958.6	4.0	14.0
SH144	65.0 (C4)	51.0 (C3)	928.8	953.9	5.4	9.7

2.2 Cyclone Impacts on Sea Ice in the Arctic during Summer

The primary data used for this study includes sea ice concentration and atmospheric fields from the European Centre for Medium Range Weather Forecasts (ECMWF) ERA-5 Reanalysis (Hersbach et al. 2020). All ERA-5 fields are available on a common 0.25-degree (~ 31 km) grid and the atmospheric fields are updated hourly, though we only use 6-hourly data in this study. Sea ice concentration (SIC) in ERA-5 is updated once daily at 00Z based on the real-time European Organisation for the Exploitation of Meteorological Satellites (EUMETSAT) Satellite Application Facility on Ocean and Sea Ice (OSI-SAF) product (Eastwood et al. 2014, Hirahara et al. 2016). The OSI-SAF product is generated from an algorithm that estimates SIC using corrected brightness temperature observations from passive microwave satellites.

The geographic focus of this study is on the marginal seas in the Pacific sector of the Arctic, defined here as the region from 70–80N and from 120E–130W. Because this region encompasses the East Siberian, Chukchi, and Beaufort Seas, we henceforth refer to it as the ECB region. We focus on this region for multiple reasons. First, the sea ice within the ECB region is frequently impacted by cyclones during the summer months that develop within a low-level baroclinic zone along the northern coast of Siberia (Crawford and Serreze 2016). In addition, this region features high interannual variability in SIC (Fig. 3A–C) and large average 5-day losses of SIC (Fig. 3D–F) during the summer months. Part of the reason for the large reductions in SIC within this region is that it encompasses a large portion of the summertime MIZ, defined by sea ice concentrations between 15% and 80% (Strong et al. 2017). The MIZ

primarily consists of young, thin, and broken sea ice that is more apt to drift in response to surface winds and ocean currents than the more consolidated pack ice. Finally, from a seasonal perspective, the sea ice cover within the ECB region is often what differentiates years with anomalously low and high pan-Arctic sea ice cover.

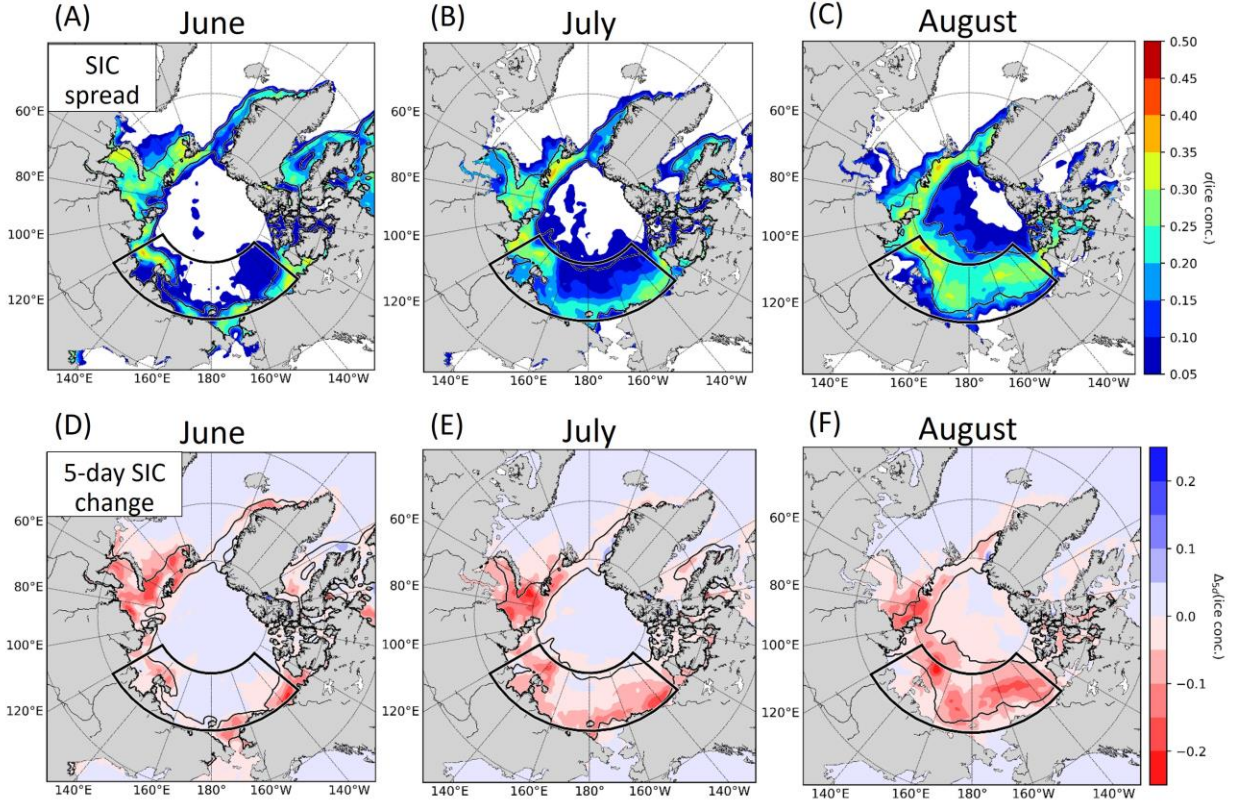


Fig. 3 — Standard deviation of sea ice concentration (SIC) over all days in June (A), July (B), and August (C), and the monthly averaged 5-day change in SIC in June (D), July (E), and August (F). The ECB region is outlined with a thick black line in each panel. The black contours are the 15% and 80% sea ice concentration contours outlining the marginal ice zone.

Although the melt season typically begins in May and ends in September (Persson 2012), we restrict our analysis to the core months of the melt season (June, July, and August) in which each day is climatologically associated with sea ice loss and a sizeable fraction of the study area is covered by the MIZ. We also only consider cyclone cases that occurred during the recent period of 1999–2018 in order to capture the effects of cyclones on the younger and thinner sea ice that has come to characterize the “New Arctic” regime (Landrum and Holland 2020).

We use a database of Arctic cyclone tracks generated from the cyclone tracking algorithm of Sprenger et al. (2017) in order to identify Arctic cyclones that occur within the ECB region during the study period. This algorithm identifies local minima in ERA-5 sea level pressure and tracks the positions of these local minima through time using the method of Wernli and Schwierz (2006, their section 2d), but using hourly ERA-5 fields rather than 6-hourly ERA-interim data. For this study, we only consider cyclones that achieve a minimum sea level pressure ≤ 995 hPa and which are tracked continuously for at least 48 h. We define cyclone days as days in which a valid cyclone is located within 300 km of the latitude-longitude box outlining the ECB region defined above. This buffer distance around the ECB region is based on the average cyclone radius from Kriegsman and Brummer (2014), and is included to account for cyclone cases whose impacts are felt within the ECB region while the cyclone is located outside of the region. The left panels of Fig. 4 show the 00Z positions of the valid cyclone cases and the

SIC averaged over the cyclone days in each month. Consistent with Crawford and Serreze (2016), there are more valid cyclone cases in July and August compared to June. Cyclones in June primarily occur in the western half of the region, while cyclones in July and August are more evenly distributed throughout the region.

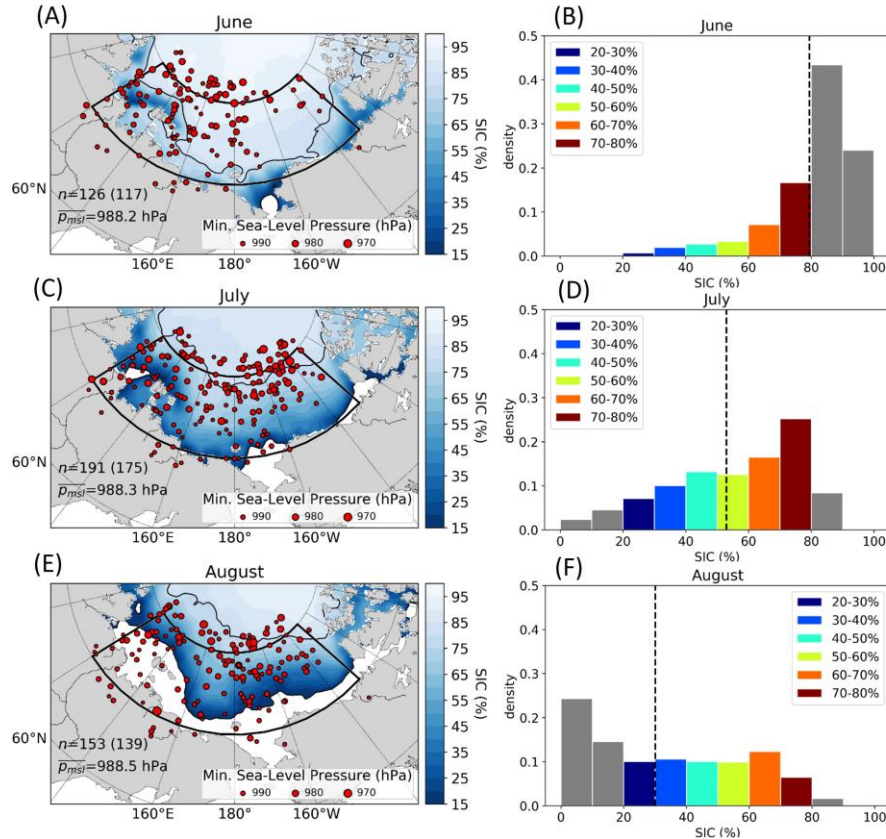


Fig. 4 — (A,C,E) cyclone cases (red dots) and the average SIC on cyclone days in each month (shading). The black contour is the 80% SIC contour denoting the poleward extent of the marginal ice zone, and the black box outlines the ECB region. The size of the dots corresponds to the cyclone minimum sea level pressure (see legend). The values printed in the lower left are the number of cyclone cases (n) and the number of unique cyclones in parentheses (as multiple cases can correspond to a single cyclone), and the average minimum sea level pressure of all cyclone cases in each month (p_{msl}); (B,D,F) probability density of monthly averaged SIC within the outlined ECB region. Bars are colored according to the SIC bins, with grey bars denoting SIC < 20% or SIC > 80%. The vertical dashed line denotes the mean SIC within the ECB region in each month.

The cyclone impacts on sea ice that can be estimated from the uncoupled ERA-5 include SIC advection by the surface winds and the surface energy flux from the atmosphere. Sea ice advection is part of the dynamic contribution of the atmosphere to local changes in SIC. We compute sea ice advection using the 10-m wind vector and the local gradient in SIC at each grid point, and assume that sea ice drifts at 2% of the surface wind speed (Spreen et al. 2011). The net surface energy flux from the atmosphere is the direct thermodynamic contribution that influences melting on top of the sea ice. We believe ERA-5 is the first reanalysis product that is suitable for this analysis because it has been shown to have smaller biases in surface winds and atmospheric surface energy fluxes in the Arctic than previous reanalysis products (Graham et al. 2019). However, ERA-5 does not contain sufficient information about the upper ocean for a detailed analysis of the indirect thermodynamic contributions of the atmosphere to sea ice loss, which involve upper ocean heating due to vertical mixing that results in lateral and basal ice melt (Steele et al. 2010, Zhang et al. 2013).

The right panels in Fig. 4 show the distribution of the monthly average SIC within the ECB region. June features predominately high-concentration sea ice with an average concentration of ~80%, while

August features predominately low-concentration sea ice with an average concentration of $\sim 30\%$. This distributional shift in SIC over the course of the melt season has important implications for the amount of sea ice loss that can be expected to occur in response to a cyclone. The net shortwave radiative flux at the surface, for example, varies strongly with SIC due to differences in the average surface albedo. Consequently, the cloud shading effect of cyclones is expected to be greater in regions of low-concentration sea ice where more shortwave radiation is absorbed rather than reflected at the surface (Perovich 2018). Furthermore, low concentration sea ice has weaker internal forces than high concentration sea ice, which allows such sea ice to drift more easily in response to the surface wind forcing (Spreen et al. 2011). Therefore, the advective effects of surface winds on sea ice will also tend to be larger in areas with low SIC. In order to account for these ice concentration dependencies in the ensuing analysis, we compute sea ice advection and atmospheric energy flux at the surface separately for grid points categorized by their SIC into six 10%-wide bins ranging from 20% to 80% ice concentration.

3. RESULTS

3.1 The Effect of Increasing Vertical Wind Shear on Tropical Cyclone Rapid Intensification

3.1.1 Intensity and structural changes due to increasing shear

Increasing shear interrupts TC intensification in each simulation and causes a more pronounced weakening of the initially stronger TCs (colored lines in Figs. 2a–c). After shear increases, it takes 11–19 h for TCs to weaken relative to the parallel simulations that maintain 3 m s^{-1} shear around the TCs. This relative weakening time is determined objectively as the time when a running 5-h average of the maximum tangential wind speed in the increasing shear simulation first becomes 1 m s^{-1} weaker than in the parallel simulation. The magnitude of deep-layer shear ranges from $10\text{--}12\text{ m s}^{-1}$ at the relative weakening time, indicating that weakening begins before the shear reaches the target value of 15 m s^{-1} . Differences in the initial intensity response are most apparent in Fig. 2b, which shows the minimum sea-level pressure in each simulation relative to when the shear starts to increase. Increasing shear prevents intensification of the weakest initial TC, which has an intensity of $\sim 12\text{ m s}^{-1}$ ($\sim 1000\text{ hPa}$) throughout the simulation. Increasing shear for the TCs in the early stages of RI (SH48 and SH60) initially causes the maximum tangential winds to weaken by $5\text{--}8\text{ m s}^{-1}$. Thereafter, the intensity fluctuates around 30 m s^{-1} (980 hPa). Increasing shear later during RI (SH72 and SH84) causes the TC to weaken by more than 10 m s^{-1} before reaching an intensity of $\sim 32\text{ m s}^{-1}$ ($\sim 970\text{ hPa}$) that remains nearly steady from 60 h after the shear starts increasing until the end of the simulation. Increasing shear at the end of RI (SH96) results in the most dramatic weakening in this set of simulations, with the maximum tangential wind speed decreasing by more than 20 m s^{-1} in the 24 h after the relative weakening time, and leveling off at a similar final intensity (33 m s^{-1}) as the TC in SH72 and SH84. In SH120 and SH144, increasing shear causes the TC to weaken to a near-steady intensity of $\sim 39\text{ m s}^{-1}$ ($\sim 955\text{ hPa}$), which is the strongest near-steady intensity in this set of simulations. These results demonstrate that the TC intensity response to increasing shear depends on the strength of the TC vortex.

Increasing shear also causes different tilt evolutions depending on the initial TC intensity. Figure 2d demonstrates this by showing differences in the 1–6 km vortex tilt magnitude between simulations with increasing shear and constant 3 m s^{-1} shear. In each of the increasing shear simulations, the tilt magnitude begins to increase relative to the TCs in constant shear from 6–12 h after the shear starts to increase, roughly 6 h before the relative weakening time. The tilt of the weakest initial TC (SH24) increases rapidly and the vortex completely decouples after 24 h. For initially stronger TCs, increasing shear causes the tilt to increase to between 10 and 30 km, and for SH120 and SH144, the tilt never exceeds 10 km during the simulations. Figure 2d also shows that the 1–6 km tilt magnitude increases more gradually as the initial intensity of the TC increases.

Differences in the initial tilt magnitudes among the simulations are broadly consistent with the different reflectivity patterns from 18–24 h after the shear increases (Fig. 5, middle row). During this time period, the reflectivity in all of the simulations has an azimuthal wavenumber-1 asymmetry maximized to the left of the shear vector. The TC in SH24 has a distinct mesoscale convective complex to the left of shear, while the TCs in SH48–SH120 maintain a more azimuthally extensive precipitation pattern with a maximum reflectivity in the upshear-left quadrant. The bottom row of Fig. 5 shows the simulated radar reflectivity averaged from 96–102 h after the shear increases, which is during the period of near-steady intensity. The precipitation in SH24 is organized in a (azimuthally) narrow band downshear. TCs in the other simulations have an open southern eyewall and expansive precipitation extending ~200 km to the left of shear. Despite the similar precipitation structures of TCs in SH48–SH144 at these later times, the vortex tilt is 10–15 km larger in SH48 and SH60 than in SH120 and SH144 (Fig. 2d).

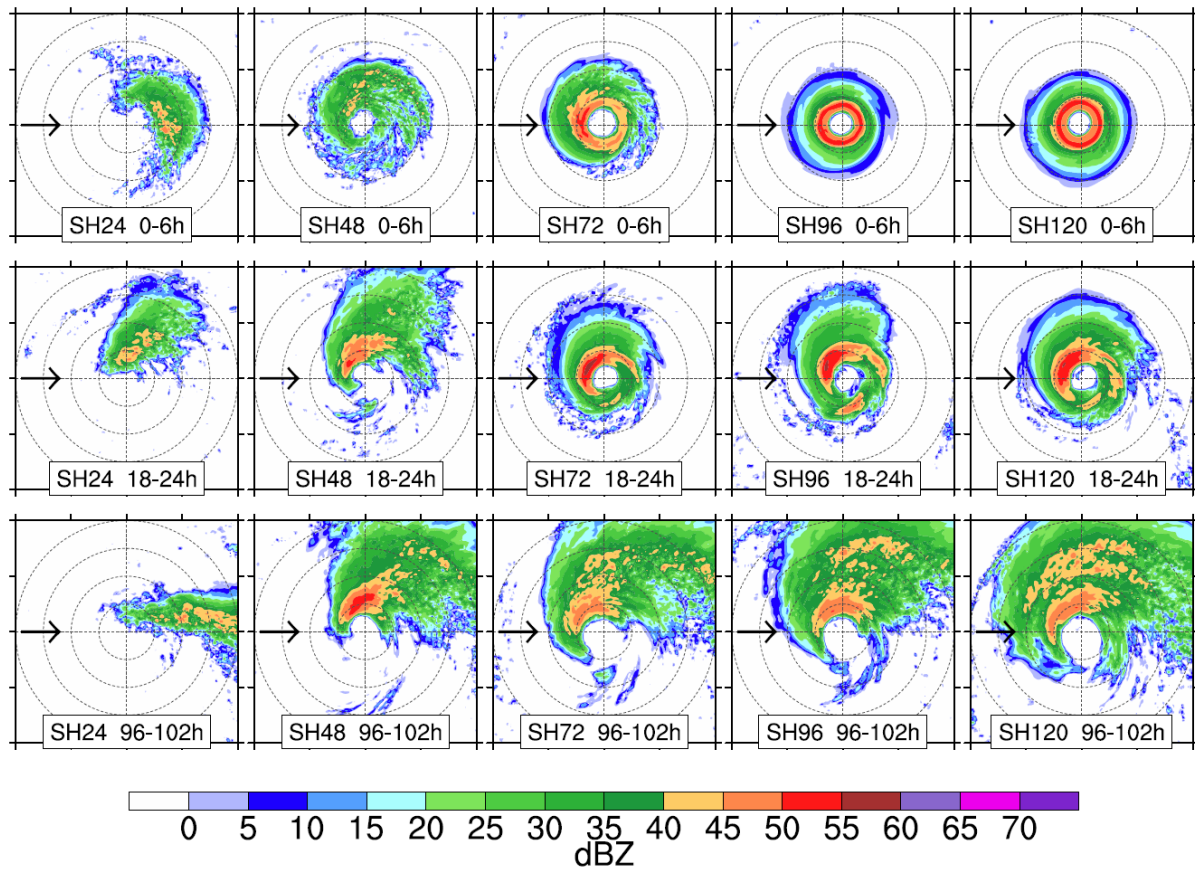


Fig. 5 — Simulated radar reflectivity at 1 km altitude in each simulation averaged from 0–6 h (top row), 18–24 h (middle row), and 96–102 h (bottom row) after shear starts increasing. Range rings are drawn every 50 km. Values printed along the top of each panel are (from left to right) the 1–6 km vortex tilt magnitude, the maximum surface (10-m) wind speed, and the radius of maximum surface winds. The black arrow on the left side of each panel indicates the direction of the large-scale vertical wind shear.

The intensity and structural evolutions described above motivate two key questions:

- (1) What causes the differences in the initial weakening response to increasing shear, and
- (2) Why do the mature initial TCs maintain a stronger near-steady intensity (and a smaller tilt magnitude) at the end of the simulations than the weaker initial TCs?

The remainder of our analysis addresses these two questions by comparing the simulations primarily during two time intervals highlighted by gray boxes in Fig. 2b, which encompass the initial weakening response to increasing shear and the period of near-steady intensity.

3.1.2 Vortex resiliency to tilt and lateral ventilation

One aspect of a TC's resiliency to vertical wind shear is the ability of the vortex to resist being tilted by the shear. Previous studies have found that the radial structure of the tangential wind field influences the ability of a sheared TC vortex to remain upright (Jones 1995, Reasor et al. 2004, Riemer and Montgomery 2013, Reasor and Montgomery 2015). Radial profiles of tangential wind speed and relative vorticity around the time when the shear starts to increase in these simulations are shown in Fig. 6 and provide a qualitative sense for how the evolving structure of the initial vortex affects the subsequent tilt response to increasing shear. The expansions of the outer wind field, in particular, appear to correspond to changes in the initial tilt evolution shown in Fig. 2d. In the four simulations spanning the early part of RI (SH48-SH84), the 34-kt wind radius indicated by diamonds in Fig. 6 increases by 34 km or 23%. The amount by which the vortex tilts over the first 36 h steadily decreases as the initial outer wind field expands (Fig. 2d). From these four simulations, however, it is difficult to tell whether the reductions in vortex tilt are due exclusively to the broadening of the outer wind field because of the simultaneous rapid spin-up of the inner core. In contrast, a more abrupt expansion of the outer wind field that takes place between the SH96 and SH144 simulations, in which the 34-kt wind radius increases by 22 km or 12%, is not accompanied by a large increase in the maximum tangential wind speed (Fig. 6). The TCs in SH120 and SH144 that are exposed to increasing shear after this second, more abrupt expansion of the outer wind field are the most resilient in this set of simulations, maintaining tilt magnitudes ≤ 10 km and Category-3 intensity despite 15 m s^{-1} of vertical wind shear. This qualitative analysis suggests that increases in both the size of the outer wind field and the intensity make TCs more resilient to tilting in response to increasing shear.

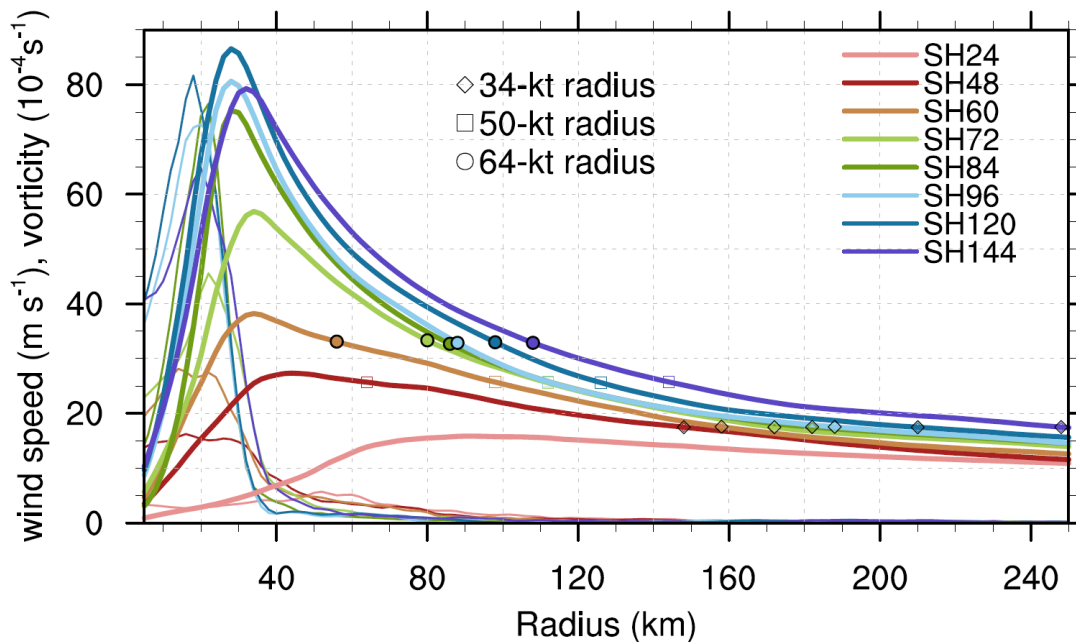


Fig. 6 — Radial profiles of azimuthally averaged tangential wind at 1 km altitude (thick lines) and relative vorticity at 2 km altitude (thin lines). Filled circles, open squares, and filled diamonds denote the 64-kt, 50-kt, and 34-kt wind radii, respectively. All quantities are averaged during the first 6 h after shear starts increasing in each simulation.

Jones (1995) argued that the coupling between vorticity anomalies at different vertical levels in a TC facilitates the cyclonic precession toward a left-of-shear orientation that prevents further increases in vortex tilt. She quantified vertical coupling using the Rossby penetration depth (R_d), which measures the

vertical extent of the induced winds above and below a vorticity anomaly (Hoskins et al. 1985). The Rossby penetration depth is expressed as $R_d = I^*L/N$, where I is the square root of the inertial stability [$I^2 = (f + \zeta)(f + 2v_t/r)$], where f is the Coriolis parameter, ζ is the relative vorticity, v_t is the tangential wind speed, and r is radius], L is the horizontal length scale of the vortex, and N is the Brunt-Väisälä frequency. Accordingly, stronger tangential winds (larger I), a broader tangential wind field (larger L), and lower static stability (smaller N) promote stronger vertical coupling within the vortex. In order to diagnose vertical coupling in the 1–6 km layer used for the vortex tilt calculations in this study, we compute R_d separately at 1 and 6 km altitudes using model output averaged during the first six hours after shear starts to increase. We define the horizontal length scale of the vortex (L) at each altitude as the outermost radius where the azimuthally averaged relative vorticity is $> 3 \times 10^{-5} \text{ s}^{-1}$. This threshold was determined empirically to include both the high vorticity in the inner core and the elevated vorticity at radii $> 100 \text{ km}$ that is generated from intense rainband convection. We apply this threshold after smoothing the vorticity profile 100 times using a 1-2-1 smoother. We then compute inertial stability at altitudes of 1 and 6 km using the tangential winds and relative vorticity radially averaged within the distance L from the TC center at each level. Finally, N is the pressure-weighted vertical average from 1–6 km of the dry Brunt-Väisälä frequency, which we compute from the temperature profile radially averaged within the distance L computed at 1 km altitude. Table 2 contains the values of R_d along with each term used in its calculation at 1 and 6 km altitudes.

In the SH24 simulation, R_d at each level is $< 5 \text{ km}$, indicating that the winds from a vorticity anomaly at 6 km do not penetrate downward to 1 km, and vice versa. Therefore, the mutual interaction within this layer of the vortex is weak and likely explains why this TC is completely decoupled by the sheared flow. The magnitude of R_d is largest for the TCs in SH120 and SH144, which is due to these TCs having both a strong and radially expansive tangential wind field and likely explains why the TCs in these two simulations tilt the least. Static stability is similar among the simulations, indicating that differences in R_d are primarily due to differences in the intensity and size of the tangential wind field. However, increases in vortex size across the set of simulations make a smaller contribution to increasing R_d than the more substantial increases in inertial stability at both 1 km and 6 km altitudes.

Table 2 — Values of the Rossby penetration depth (R_d) at 1 and 6 km altitudes (left two columns) and the values of inertial stability (I), vortex size (L), and inner core static stability (N) used in the calculation of R_d at each level (see text for details).

	R_d at 1 km (km)	R_d at 6 km (km)	I at 1 km ($\times 10^{-4} \text{ s}^{-1}$)	I at 6 km ($\times 10^{-4} \text{ s}^{-1}$)	L at 1 km (km)	L at 6 km (km)	N ($\times 10^{-2} \text{ s}^{-1}$)
SH24	4.7	3.7	3.3	2.7	213	142	1.24
SH48	8.4	6.9	6.5	6.1	163	142	1.25
SH60	10.2	8.3	10.1	7.0	131	152	1.29
SH72	14.1	11.1	8.0	8.5	222	163	1.25
SH84	15.6	12.6	9.5	8.5	211	187	1.26
SH96	18.6	13.2	10.7	13.3	218	124	1.24
SH120	19.3	14.3	14.8	13.8	186	132	1.28
SH144	19.7	16.0	11.0	10.1	225	199	1.26

Resiliency to vertical wind shear can also be understood in terms of lateral intrusions of environmental air into the TC inner core (i.e., lateral ventilation). Tang and Emanuel (2010) demonstrated that the equilibrium intensity of an axisymmetric TC subjected to lateral ventilation is inversely related to the lateral ventilation rate. For the simulations in the present study, this result implies that, despite experiencing the same environmental vertical wind shear, the stronger initial TCs must experience less lateral ventilation than the weaker initial TCs in order to ultimately reach a higher near-steady intensity (Fig. 2b). One possible explanation for this is that a stronger and broader tangential wind field reduces

how much a given magnitude of vertical wind shear is capable of ventilating a TC (e.g., Riemer and Montgomery 2011). To examine this further, Fig. 7 plots the storm-relative zonal wind speed versus the TC circulation at each 1-km height interval between 1 and 12 km altitude in each simulation. Quantities are time averaged from 18–24 h after the shear starts to increase. The circulation is computed by averaging relative vorticity within 50 km (“eyewall radius”, Fig. 7a) and 150 km (“rainband radius”, Fig. 7b) of the TC center at each level. Gray shaded areas highlight parameter values within this phase space that Riemer and Montgomery (2011) found to be associated with significant inflow rates of environmental air (see their Fig. 9). At the eyewall radius (Fig. 7a), the altitude above which lateral ventilation is likely to occur based on these parameters increases from around 6 km in SH24 to 11 km in SH144. This implies that, for the sheared wind profile used in these simulations, the more mature TCs are increasingly isolated from destructive inflows of ambient air in the middle troposphere. We find similar results at the rainband radius (Fig. 7b), except the minimum altitude of lateral ventilation is 1–2 km lower than at the eyewall radius. Although this analysis does not account for the development of a divergent outflow, which we will show reduces lateral ventilation at higher altitudes, it demonstrates how the deepening circulation in more mature TCs effectively seals off the inner core from lateral ventilation through a deeper layer of the middle troposphere.

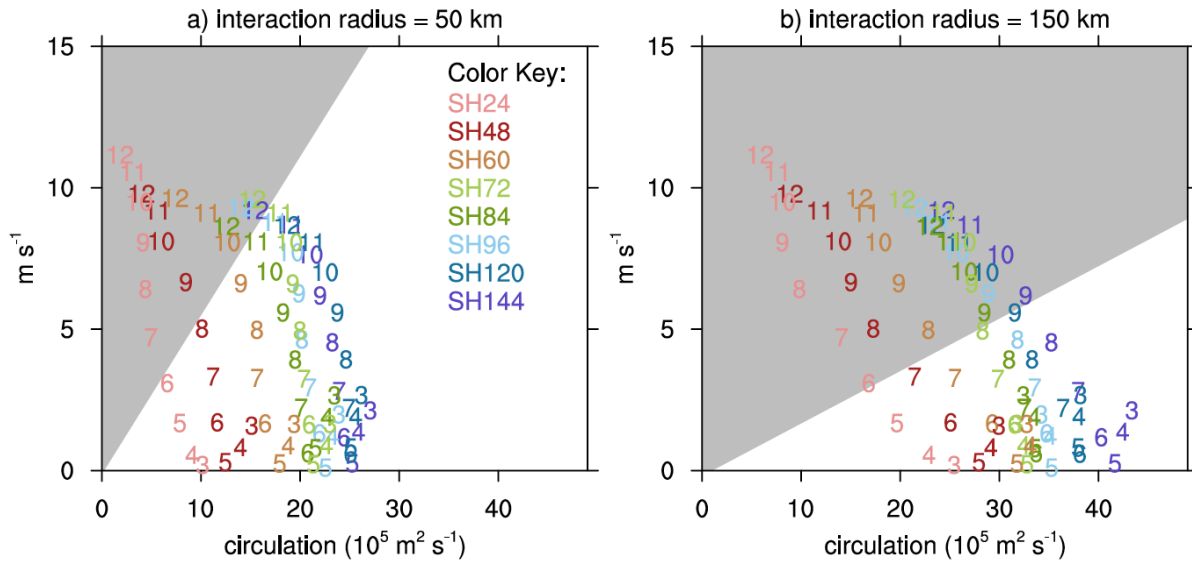


Fig. 7 — Scatter plot of storm-relative zonal wind speed (y-axis) and circulation computed within a radius of 50 km (a), and 150 km (b) of the TC center (x-axis). Values are averaged from 18–24 h after shear starts increasing in each simulation. Gray shading indicates parameter values within the same 2-d phase space for which Riemer and Montgomery (2011) found significant lateral ventilation at each radius in their kinematic model.

To quantify how much mid-tropospheric air is actually transported laterally into the TC inner core, we compute backward trajectories from the inner core using the Lagrangian Trajectory Tool (LAGRANTO, Sprenger and Wernli 2015). Backward trajectories are computed on the static 18-km WRF grid with 12-minute output. Trajectories are seeded within a 100 km radius of the TC surface center position every 500 m between 5–16 km altitude. We classify a trajectory as a lateral transport trajectory if it originates at a radius > 100 km from the TC center and travels horizontally from its origin to the seed location in one of three vertical layers: 5–8 km, 8–11 km, and 11–14 km. We enforce the horizontal transport criterion by requiring trajectory altitudes to remain within ± 2 km of the upper and lower boundaries of the seed layers at all times. Backward trajectory calculations are carried out in two 12-h time intervals: the first interval starts 30 h and ends 18 h after the shear starts increasing and captures the initial TC response to increasing shear; the second interval starts 108 h and ends 96 h after the shear starts increasing, during the near-steady intensity period. We note that trajectories often follow elliptical paths

around the storm and will periodically approach the TC center. This fairly simple analysis classifies such trajectories as lateral transport trajectories even if they are not entrained in inner-core convection.

Figure 8 shows the percentage of trajectories seeded in each layer at 30 h (Fig. 8a) and 108 h (Fig. 8b) that are classified as lateral transport trajectories. Shortly after the shear increases (Fig. 8a), more than 60% of the trajectories in all three layers are laterally transported from the environment in SH24. As the TC exposed to increasing shear matures, these percentages decrease to less than 20% from 11–14 km due in large part to the development of a divergent outflow, which deflects inflowing air parcels at these altitudes (e.g., Ryglicki et al 2019). The percentage of lateral transport trajectories also decreases in the 5–8 km and 8–11 km layers, but remains above 50% in the 8–11 km layer in SH48–SH96. The 8–11 km layer experiences the most lateral transport because it is located just beneath the outflow layer but at a high enough altitude to experience storm-relative flow from the increasing shear that is stronger than in the 5–8 km layer (Figs. 1a, 7). The black dots in Fig. 8 denote the percentage of lateral transport trajectories in the parallel simulations with constant 3 m s^{-1} shear. Even in weak shear, a considerable fraction of the trajectories is laterally transported toward the inner core. However, increasing shear to 15 m s^{-1} generally results in 10–25% more lateral transport trajectories in most simulations in the 8–11 km layer. In cases like SH84, where the percentage of lateral transport trajectories in the 8–11 km layer is similar in the parallel (constant shear) and increasing shear simulations, the mean properties of these trajectories are quite different; the average equivalent potential temperature (θ_e) at the origin position is 344 K in the increasing shear simulation and 348 K in the constant shear simulation, and the average origin distance of trajectories is 378 km in the increasing shear simulation versus only 146 km in the constant shear simulation. Therefore, the percentages in Fig. 8 do not give a full sense of the destructive potential of lateral transport trajectories, which is why we will later attempt to estimate the thermodynamic impacts of lateral transport.

During the period of near-steady intensity (Fig. 8b), the TCs in all simulations except SH24 experience less lateral transport between 5–11 km than earlier in the simulations (cf. Fig. 8a). Note that the TC in SH24 has been completely sheared apart by this time, so nearly all trajectories originating above the surface center are laterally transported to their seed locations. Lateral transport trajectories in all three layers make up less than 20% of all trajectories in SH84–SH144, but are even less frequent in the parallel simulations with weak shear.

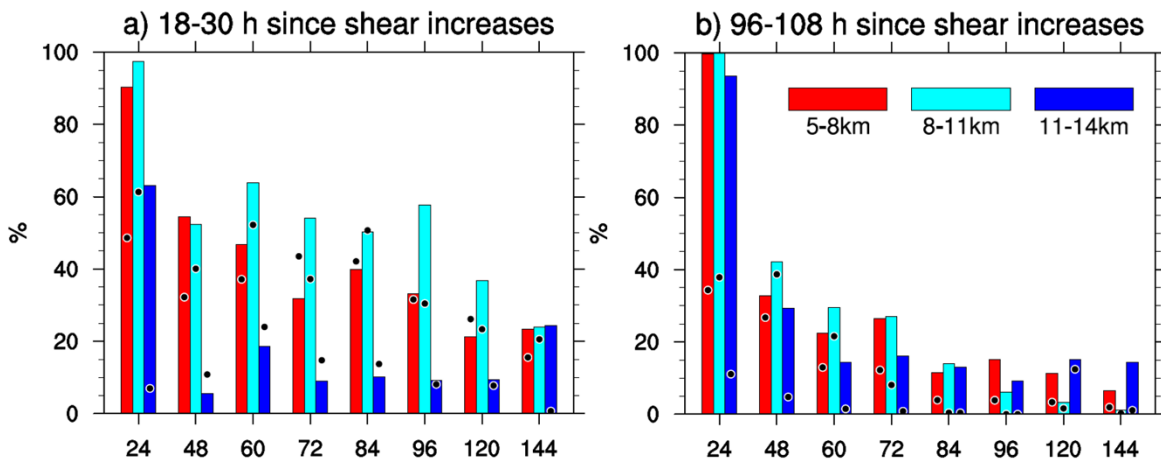


Fig. 8 — Percentage of backward trajectories seeded within 100 km of the TC center and at 30 h (a) and 108 h (b) after shear starts increasing which are laterally transported from the surrounding environment to the inner core in the 5–8 km (red), 8–11 km (cyan), and 11–14 km (blue) layers. Black dots denote the percentages from the parallel simulations with constant 3-m s^{-1} shear. Values along the x-axis indicate the hour at which shear starts to increase in each depicted experiment.

Figure 9 reveals the structure of lateral transport in azimuth-height cross-sections of θ_e averaged between 50–100 km from the TC center in four selected simulations. From 18–24 h after shear starts

increasing (Fig. 9a-d), lower- θ_e air encroaches upon the southern and western sides of the TC in SH48 between 6–9 km altitude (Fig. 9a). This mid-level θ_e reduction upshear becomes less pronounced in stronger initial TCs. A second, weaker signature of dry air encroachment is evident on the northwest (upshear-left) side of the TCs between 1–4 km altitude. Both signatures of reduced θ_e air qualitatively agree with the patterns of radial ventilation documented by Alland et al. (2021b, their Fig. 5), which they found to be a result of easterly storm-relative inflow in the lower levels, and vortex tilt in the mid-to-upper levels. Low- θ_e air encroaches downwind of the tilted convective updrafts in each simulation, such that precipitation falling into this dry air likely contributes to the formation of the mesoscale downdrafts shown as black contours in Fig. 9. During the period of near-steady intensity (Fig. 9e-h), low- θ_e air encroaches between 3–8 km altitude in the upshear-right quadrant. Consistent with the frequencies of lateral transport trajectories between 5–8 km shown in Fig 8b, the amount of θ_e reduction within this layer decreases as the near-steady intensity of TCs increases.

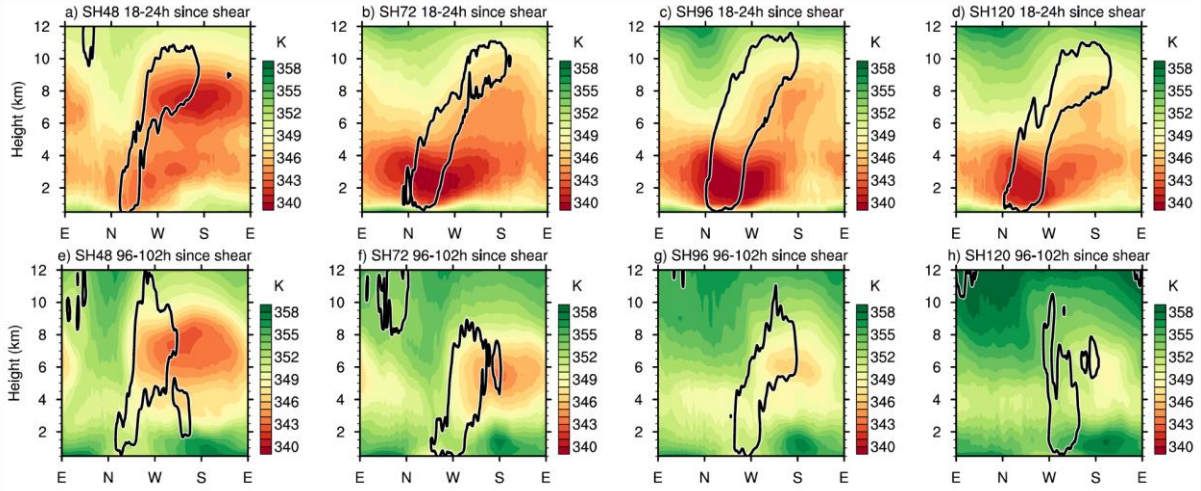


Fig. 9 — Azimuth-height cross-sections of θ_e (shading) and vertical velocity $\leq -0.2 \text{ m s}^{-1}$ (thick black contour) averaged from 50–100 km from the TC center in SH48, SH72, SH96, and SH120 (left to right). Fields are time-averaged from 18–24 h (a-d) and 96–102 h (e-h) after shear starts increasing. Cardinal directions looking outward from the TC center appear along the x-axis.

We can use the thermodynamic properties of the lateral transport trajectories to estimate their impact on inner core θ_e . For this analysis, we utilize a modified version of Eq. 2 from Cram et al (2007):

$$\Delta\theta_e = P_L \left(\langle \theta_e(t_s - 3h) - \theta_e(t_s) \rangle + \langle \theta_e(t_s) - \bar{\theta}_e(t_s) \rangle \right), \quad (2)$$

where $\Delta\theta_e$ is the change in azimuthally averaged inner-core θ_e due to lateral ventilation, P_L is the percentage of trajectories in each layer classified as lateral transport trajectories, and angled brackets denote averages over all lateral transport trajectories in each layer. This equation accounts for the change in θ_e along backward trajectories during the 3 h prior to their seed time, t_s (first term in brackets on the right hand side of Eq. 2), which represents energy air parcels extract from the inner-core environment, and the θ_e deficit of trajectories relative to the azimuthally averaged θ_e of the inner core at the seed position (second term in brackets on the right hand side of Eq. 2). Note that the air parcels within each layer of the inner core that arrived by means other than lateral transport also induce a change in inner-core θ_e that is not accounted for in Eq. 2.

Figure 10 shows the $\Delta\theta_e$ values computed from the early (a) and late (b) sets of trajectories and in the three vertical layers from Fig. 8. During the initial weakening period (Fig. 10a), lateral ventilation in both the 5–8 km and 8–11 km layers actually increases with increasing intensity of the initial TC from SH24 to SH84, and then decreases from SH96 to SH144. Consequently, the magnitude of θ_e reductions increases from about 0.5 K in SH24 to 1.4 K in SH84 in the 5–8 km layer, and from about 0.2 K in SH24 to 1.8 K in SH84 in the 8–11 km layer. One explanation for the greater amount of cooling due to lateral

transport with increasing storm intensity from SH24 to SH84 is that the radial gradients in mid-tropospheric θ_e just outside the inner core strengthen as the TC rapidly intensifies. The resultant cooling of the inner core just below the outflow layer by nearly 2 K in the fairly intense TCs in SH60–SH96 likely plays a role in the significant weakening that occurs in response to increasing shear. By contrast, the θ_e reductions due to lateral ventilation in the upper troposphere from 11–14 km are much smaller in magnitude than in the middle troposphere and do not vary systematically with the initial intensity or size of the TC. As discussed in Tang and Emanuel (2012), lateral ventilation tends to be weaker in the upper troposphere due to the weak radial gradients in θ_e . Furthermore, Ryglicki et al. (2019) noted that many inflowing air parcels are deflected around the inner core at these altitudes when there is a component of the divergent outflow directed upshear (which is the case in all simulations except SH24). This strongly limits the potential for lateral ventilation to cool the inner core between 11–14 km in our simulations.

During the period of near-steady intensity (Fig. 10b), reductions of inner-core θ_e due to lateral ventilation are smaller than when the shear first begins to increase. This is particularly evident in the 8–11 km layer, where lateral ventilation reduces θ_e by no more than 1 K in any of the experiments (we disregard the SH24 simulation during this period because the vortex is completely sheared apart by this point and the θ_e reductions in Fig. 10b become difficult to interpret). Unlike the initial weakening period in Fig. 10a, the θ_e reductions between 5–8 km and 8–11 km steadily decrease for the stronger and larger TCs, with small θ_e reductions in all three layers in SH144. This is consistent with the expectation from Fig. 7 that larger and stronger TCs are more kinematically insulated from storm-relative inflow between 5–11 km than smaller and weaker TCs. These results further suggest that lateral ventilation influences the near-steady intensity of TCs in these simulations.

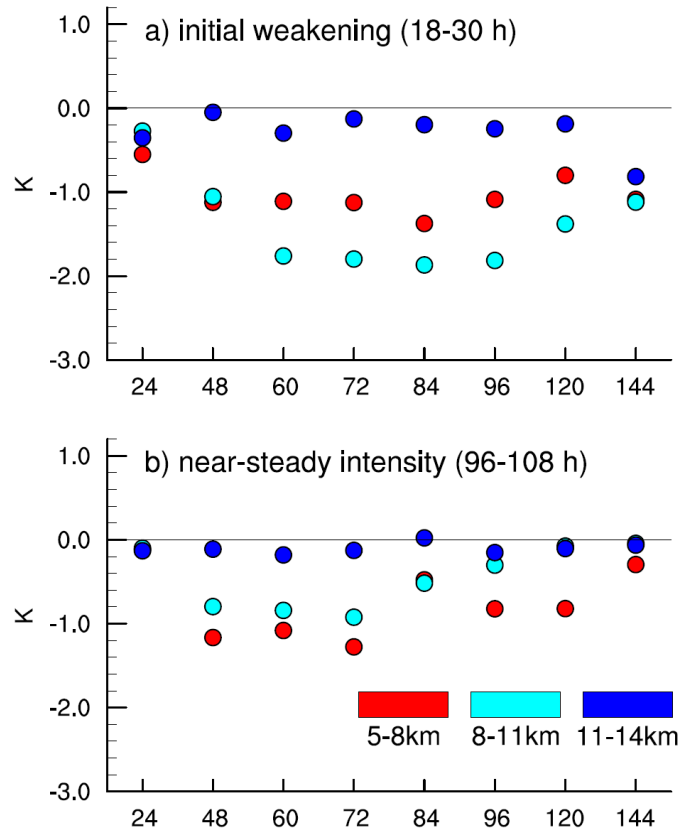


Fig. 10 — the average amount by which lateral ventilation changes axisymmetric θ_e within 100 km of the TC center and in each vertical layer shown in the legend estimated from backward trajectories initialized at (a) 30 h and (b) 108 h after the shear starts increasing. Values are computed based on Eq. 2, which is adapted from Cram et al. (2007). As in Fig. 8, values along the x-axis indicate the hour at which shear starts to increase in each depicted experiment.

3.1.3 Thermodynamic Modification of Boundary Layer and the Inner Core

As a TC tilts in response to increasing vertical wind shear, downdrafts begin to flush low-entropy air into the boundary layer, which is swept inward toward the TC center (Riemer et al. 2010). This low-entropy air can feed into eyewall updrafts, decreasing the amount of work the TC does against frictional dissipation and thereby weakening the storm. Boundary layer flushing is evident in the θ_e fields shown in Fig. 11 averaged from 0–6 h (top row), 18–24 h (middle row), and 96–102 h (bottom row) after the shear starts to increase. The areas of reduced θ_e from 0–6 h (Fig. 11, top row) also appear in the parallel simulations with 3-m s^{-1} shear (not shown), which likely means that the shear at this early stage is not yet strong enough to enhance boundary layer flushing. However, by 18–24 h (Fig. 11, middle row), the θ_e reductions in the upshear quadrants of the increasing shear simulations are larger than in the parallel simulations, indicating enhanced flushing due to the increasing shear. The low- θ_e air arrives in the boundary layer via downdrafts located 30–150 km from the storm center in the upshear-left (USL) quadrant (black contours). Once the TCs reach near-steady intensity (Fig. 11, bottom row), a band of downdrafts emanates from the USL quadrant to a radius of more than 200 km in the downshear-left (DSL) quadrant in all of the simulations except SH24. These downdrafts are located along the leading edge of the asymmetric convection (cf. Fig. 5) and flush low- θ_e air throughout the upshear semicircle.

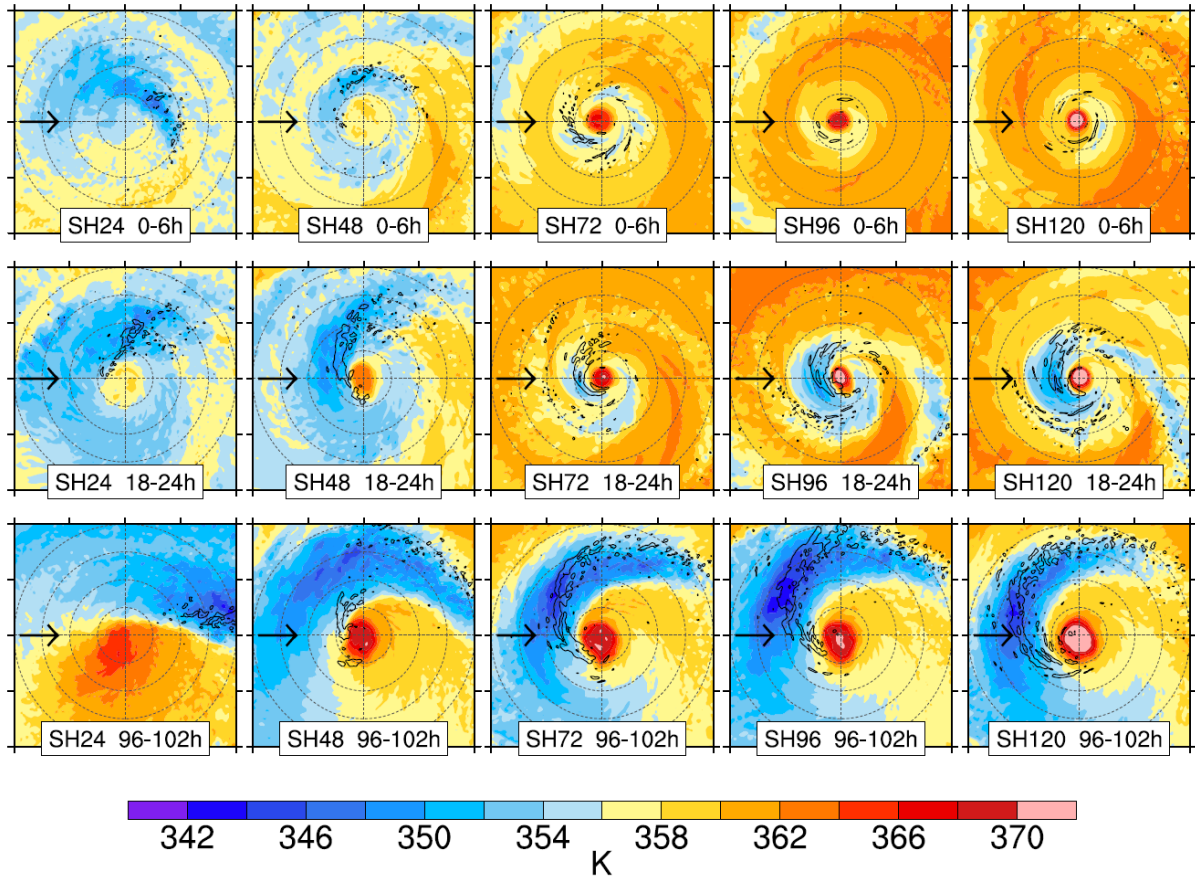


Fig. 11 — As in Fig. 5, but for θ_e averaged over the 3 lowest model levels (shading) and vertical velocity at 1.5 km altitude $\leq -0.5\text{ m s}^{-1}$ (black contours).

We quantify boundary layer flushing as the downward flux of θ_e anomalies at the top of the boundary layer, or $w_1 \theta'_e$ (Riemer et al. 2010). The prime denotes the anomaly relative to the azimuthal

mean and $w_{\downarrow}$ is the downward-directed vertical velocity (such that $w_{\downarrow}\theta'_e > 0$ denotes a downward flux of negative θ_e anomalies). Both $w_{\downarrow}$ and θ'_e are defined at an altitude of 1.5 km and are averaged in the USL quadrant in a band extending 100 km radially outward from the 10-m RMW (repeating this analysis by averaging quantities in a fixed band that extends 30–130 km from the TC center does not meaningfully change the results). Figure 12a shows times series of $w_{\downarrow}\theta'_e$ in each simulation relative to when the shear starts to increase. During the first 18 h of each simulation, downward θ_e fluxes increase to between 1–3 K m s⁻¹. The largest downward θ_e fluxes occur in the TCs exposed to increasing shear during RI (SH72–SH96). After 18 h, downward fluxes remain between 1–3 K m s⁻¹, with occasionally larger values during strong flushing episodes.

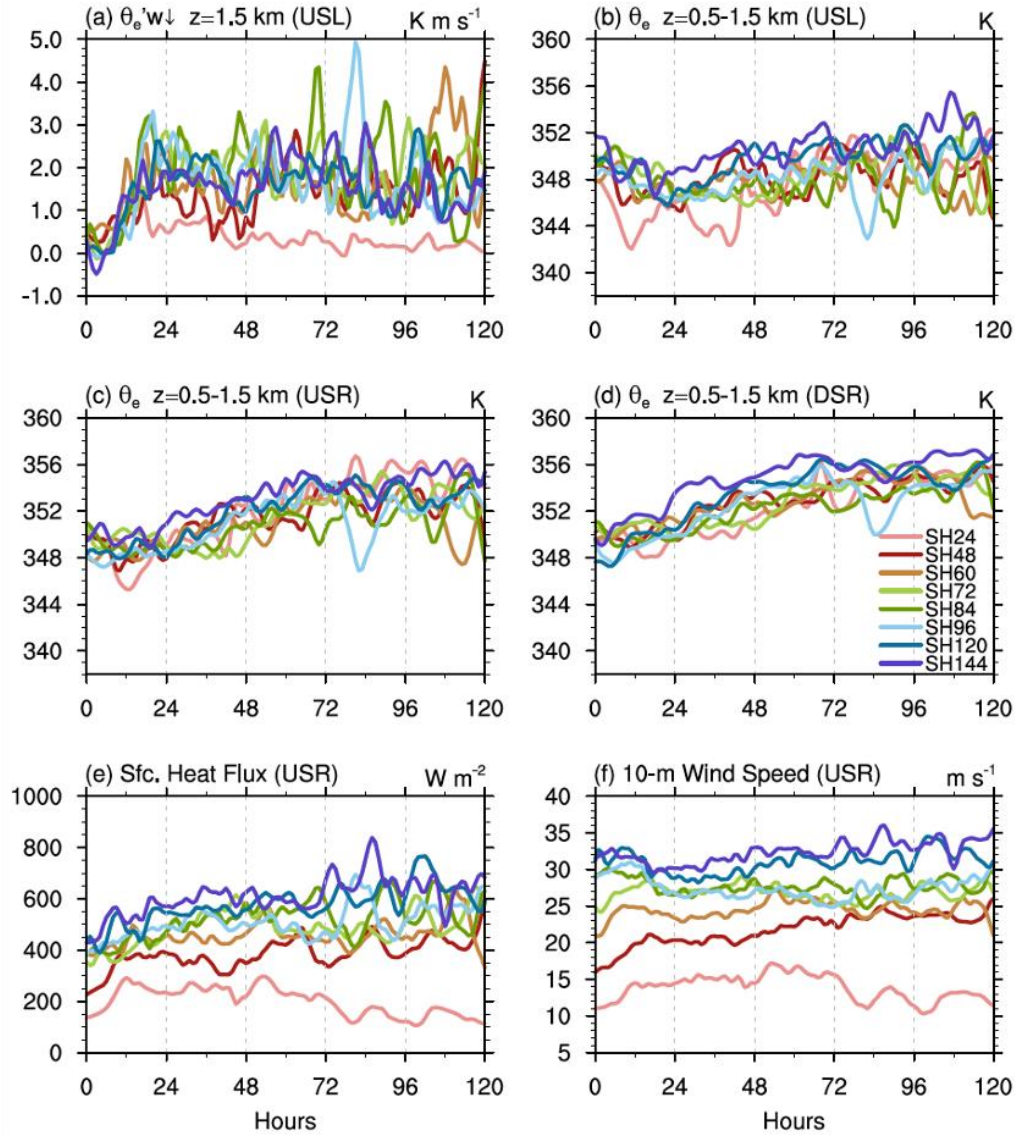


Fig. 12 — Time series of (a) downward flux of θ_e anomalies ($w_{\downarrow}\theta'_e$) in the USL quadrant, (b) θ_e in the USL quadrant, (c) θ_e in the USR quadrant, (d) θ_e in the DSR quadrant, (e) the sum of latent and sensible heat fluxes in the USR quadrant, and (f) 10-m wind speed in the USR quadrant. All quantities are radially averaged in a band extending 100 km outward from the RMW at the surface.

Figures 12b–d show θ_e averaged in the same radial band in the USL, upshear-right (USR), and downshear-right (DSR) quadrants, respectively. The average θ_e at 24 h into each simulation increases by

about 5 K moving cyclonically around the storm from the USL to the DSR quadrant, indicating that downdraft air recovers via surface enthalpy fluxes primarily within the USR quadrant. This substantial θ_e recovery within the USR quadrant is also evident during the period of near-steady intensity (Fig. 11, bottom row). Figures 12e and 12f show the average surface enthalpy flux and 10-m wind speed, respectively, in the USR quadrant. Surface fluxes are larger in the TCs that have broader and stronger tangential winds within the USR quadrant. This enables a more rapid increase in θ_e via surface enthalpy fluxes. As a result, the intense TCs in SH72–SH96 that initially have the largest downward flux of low- θ_e air into the boundary layer USL (Fig. 12a) end up having similar θ_e values in the DSR quadrant as weaker TCs with weaker flushing (Fig. 12d).

We further examine this boundary layer recovery process in a Lagrangian framework using the two sets of backward trajectories described in Section 3.1.2. The backward trajectories for this analysis are seeded every 500 m between 5–16 km altitude and within 100 km of the TC center. Instead of examining trajectories transported laterally from the middle troposphere as we did in Section 3.1.2, we now consider trajectories that reach their seed positions via moist convective updrafts from the boundary layer. Specifically, we select trajectories that originate below 1 km altitude, and have a segment that intersects the 800- and 500-hPa pressure levels in which the altitude increases by at least 3 km in 3 h and in which the maximum relative humidity is at least 95%. We focus on the boundary layer segment of the selected trajectories prior to their ascent, which begins where the trajectory has its minimum θ_e value within 1 km of the sea surface and ends where the trajectory first begins to satisfy the updraft criteria above, or where the trajectory altitude first exceeds 1 km (whichever comes first). In order to constrain the sample of trajectories to those that originate in cold pools within the boundary layer, we only consider trajectories whose minimum near-surface θ_e is in the lower quartile of all the updraft trajectories. Figure 13 shows these boundary layer segments in storm-relative space colored by θ_e for four selected simulations. For trajectories seeded at 30 h (Fig. 13a–d), the boundary layer segments begin primarily in the USL quadrant in SH48 and in the USR quadrant in SH72 and SH96. The segments are distributed more evenly among all quadrants in SH120. For trajectories seeded at 108 h (Fig. 13e–h), boundary layer segments begin to the left of shear in all simulations, consistent with where significant flushing occurs (Fig. 11, bottom row).

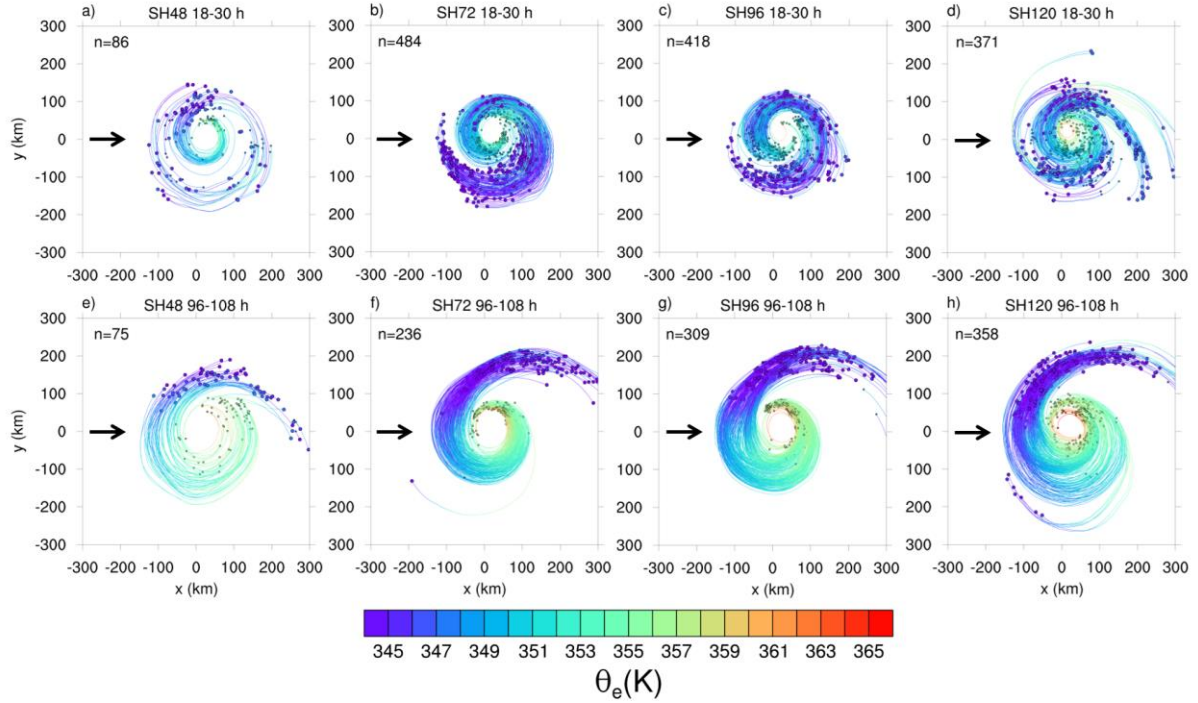


Fig. 13 — Segments of trajectories within the boundary layer ($z < 1$ km) prior to entering eyewall updrafts colored according to θ_e and plotted within TC-center relative space. Trajectories are seeded at 30 h (a–d) and 96 h (e–h) after shear starts increasing in SH48, SH72, SH96, and SH120 (from left to right). Large dots denote the minimum θ_e positions at the beginning of the boundary-layer segments. Smaller black dots mark the end of the boundary layer segments, where the trajectory begins satisfying the updraft criteria in the text or the trajectory altitude first exceeds 1 km (whichever comes first). The black arrow on the left side of each panel indicates the direction of the large-scale vertical wind shear.

Figure 14 shows distributions of the change in θ_e along the boundary layer segments of the updraft trajectories (a,d) and the θ_e at the beginning (b,e) and end (c,f) of the boundary layer segments shown in Fig. 13. For the trajectories seeded 30 h after the shear starts to increase (Fig. 14a–c), the average increase in θ_e along the boundary layer segment ranges from 3 K in SH24 to 7 K in SH144 and is up to 50% smaller than in the parallel simulations with 3 m s^{-1} shear (open black circles in Fig. 14a). TCs in the simulations in which shear increases toward the end of RI have the lowest θ_e at the beginning of the boundary layer segment (Fig. 14b), which is consistent with the stronger downward flux of low- θ_e air in these simulations (Fig. 12a). The average θ_e at the end of the boundary layer segment (Fig. 14c) indicates the entropy of air parcels when they begin ascending in convective updrafts near the TC center; lower average θ_e values at this point along the trajectories likely signify more diluted updrafts as a result of the incomplete recovery of downdraft air in the boundary layer. At this early time in the simulations, the average θ_e at the end of the boundary layer segment is lowest (349 K) in SH84 and SH96, and highest (354 K) in SH144. The lower θ_e of air parcels ascending from the boundary layer in SH96 may further contribute to the remarkable weakening that occurs in that simulation. Note from Fig. 14c that even in the parallel simulations with 3 m s^{-1} shear, the average θ_e at the end of the boundary layer segment is relatively lower between 72–96 h than at other times, suggesting that the TC is particularly vulnerable to an increase in vertical wind shear near the end of RI.

During the period of near-steady intensity, the increase in θ_e along the boundary layer segment is frequently more than twice as large as in the trajectories seeded earlier in the simulations (Fig. 14d vs. 14a). The larger increase in θ_e along the boundary layer segments later in the simulations is partly due to the minimum θ_e occurring further from the TC center (Fig. 13e–h), such that trajectories travel more than twice as far within the boundary layer before ascending in updrafts (not shown). The average θ_e at the

end of the boundary layer segment (Fig. 14f) is remarkably similar among the simulations, regardless of the near-steady intensity. For example, the TCs in SH60 and SH96 have the same average θ_e at the end of the boundary layer segments (357 K), yet the near-steady intensity of the TC in SH96 is about 10 hPa stronger than in SH60. Therefore, the thermodynamic recovery of downdraft air in the boundary layer does not appear to influence the near-steady intensity of TCs in these simulations.

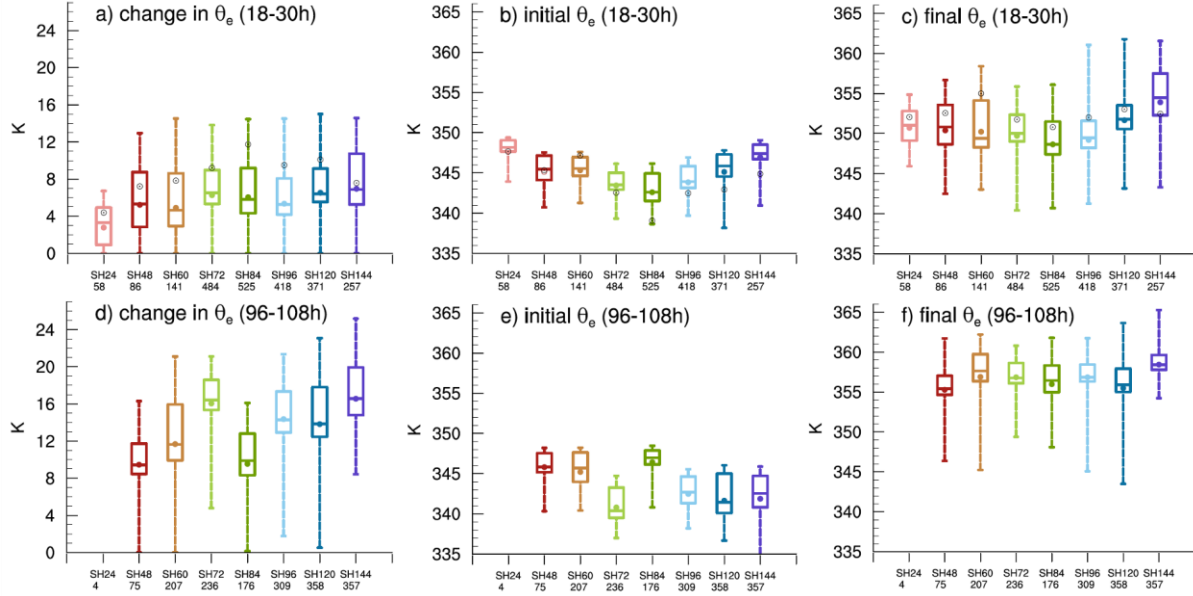


Fig. 14 — Distributions of (a,d) the change in θ_e along the boundary layer segments of updraft trajectories shown in Fig. 11, (b,e) the minimum θ_e at the beginning of the boundary layer segments, and (c,f) the θ_e at the end of the boundary layer segments. Boxes extend from the 25th to the 75th percentile, lines and dots within the boxes represent the median and mean, respectively, and whiskers extend to the extreme values. In a-c, dots are filled when the mean in the increasing shear simulations differs from that in the parallel simulations (indicated by open black circles) at the 98% confidence level. Colors follow the convention in Fig. 2. Values along the x-axis denote the number of trajectories in each distribution. Box plots are only drawn if there are at least 10 valid trajectories.

To summarize this analysis, we find that boundary layer flushing is strongest in the simulations with increasing shear toward the end of RI, and that there is a correspondence between the θ_e of air parcels ascending from the boundary layer and the amount of weakening that occurs in response to increasing shear. However, there is less of a correspondence between the θ_e of air parcels ascending from the boundary layer and the different near-steady intensities among the simulations, suggesting that boundary layer thermodynamics do not determine the near-steady intensities of TCs in these simulations.

Ultimately, the effect of increasing shear and the corresponding ventilation is reflected in the inner-core diabatic heating of the simulated TCs. The top row of Fig. 15 shows radius-height sections of the azimuthally averaged diabatic heating rate (shading) and simulated radar reflectivity (gray contours) in the four simulations shown in Figs. 9 and 13. These quantities are time-averaged from 18–24 h after the shear starts increasing. In order to demonstrate how diabatic heating changes as the shear increases, the bottom row shows differences in time-averaged heating rates from 18–24 h and 0–6 h after the shear starts increasing. Increasing shear initially shifts the largest diabatic heating rates radially away from the TC center. This is seen most clearly in the differences in SH96 and SH120 (Fig. 15, bottom row), which are negative inside and positive outside of the RMW (thick dashed line). While this outward shift is not as obvious in SH48 and SH72, the sum of the diabatic heating inside the RMW averaged from 18–24 h decreases by 30% and 45%, respectively, relative to the heating averaged from 0–6 h after shear starts increasing.

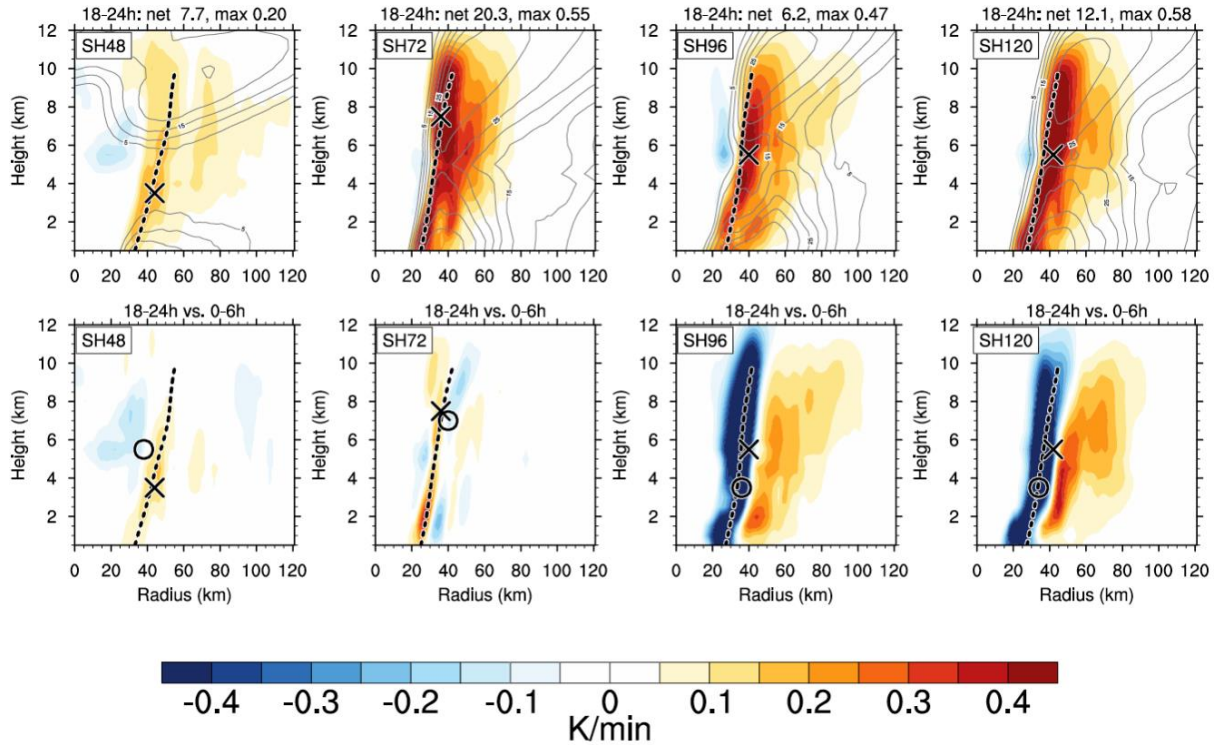


Fig. 15 — (top row) Radius-height cross sections of azimuthally averaged diabatic heating rates (shading) and simulated radar reflectivity (gray contour every 5 dBZ from 5 to 40 dBZ) in SH48, SH72, SH96, and SH120 (from left to right) time averaged from 18–24 h after shear starts increasing. The thick dashed contour is the RMW and the “X” denotes the location of the maximum heating rate. Values printed atop each panel are the sum of the heating rates radially inward of the depicted RMW (“net”) and the maximum heating rate at the location of the “X” in K min^{-1} (“max”). (Bottom row) differences between the diabatic heating rates shown in the top row and those averaged from 0–6 h after shear starts increasing. “X” and the thick dashed line are the same as in the top row, and “O” denotes the location of the maximum heating rate in the 0–6 h time interval.

Changes in diabatic heating inside the RMW during the first 24 h after shear starts increasing are highly correlated with subsequent intensity changes. Figure 16 shows the magnitude of the correlation between the percent reduction in diabatic heating inside the RMW in all eight simulations and the change in TC intensity over the same length of time but lagged by 0 to 18 h relative to the diabatic heating. At lag-0, the correlation is 0.83 for minimum sea-level pressure and 0.9 for maximum tangential wind speed. The correlation reaches a maximum (0.94) when the intensity change is computed in a time interval that lags the percent reduction in diabatic heating by 5–8 h, suggesting that the reduction in inner-core diabatic heating determines the amount of weakening that occurs in response to increasing shear, rather than the other way around. These results are broadly consistent with previous studies showing that the amount of diabatic heating inside the RMW strongly influences TC intensity change (Shapiro and Willoughby 1982, Pendergrass and Willoughby 2009, Rogers et al. 2013). These changes in diabatic heating inside the RMW may be related to vortex tilt, considering the increase in vortex tilt and the reduction in diabatic heating both precede the onset of weakening by about 6 h (Fig. 2).

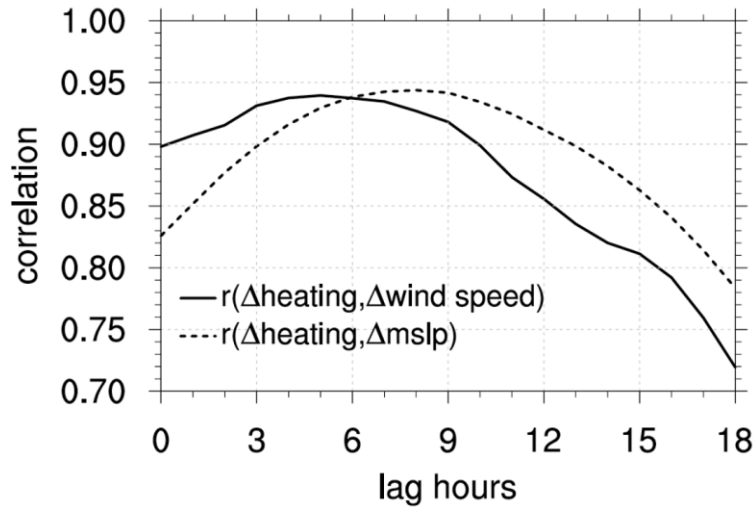


Fig. 16 — Magnitude of the correlation between the percent change in the total diabatic heating inside the RMW ($\Delta\text{heating}$, computed as the difference between the 0–6 h and 18–24 h time averages) and the change in the maximum tangential surface wind speed ($\Delta\text{wind speed}$, solid line) or minimum sea-level pressure (Δmslp , dashed line) over time intervals lagged 0 to 18 h from the time interval used to compute the change in diabatic heating. TC intensity changes at each lag are computed as differences between two 6-h time averages spaced 18 h apart (e.g. for lag 0: $\text{mslp}(18\text{--}24\text{ h}) - \text{mslp}(0\text{--}6\text{ h})$, for lag 1: $\text{mslp}(19\text{--}25\text{ h}) - \text{mslp}(1\text{--}7\text{ h})$, and so forth).

Figure 17 shows the azimuthally averaged diabatic heating rates during the period of near-steady intensity. TCs that maintain a higher near-steady intensity have a larger projection of diabatic heating onto the azimuthal mean. In particular, the stronger TCs have higher heating rates within a narrow region just inside the RMW, where heating has a large positive tendency on the tangential winds (Pendergrass and Willoughby 2009). Consequently, the integrated heating rate inside the RMW is well correlated ($r = 0.9$) with the near-steady intensity of the TCs (we exclude the SH24 simulation from the sample because the TC has been completely sheared apart by this point). It is difficult to tell whether the different diabatic heating rates at the end of the simulations are the cause or the effect of the different near-steady intensities, but if we assume that the processes acting to spin down the TC are nearly constant during this period, then it is conceivable that a configuration of diabatic heating that induces stronger positive tangential wind tendencies would allow a TC to reach a higher equilibrium intensity.

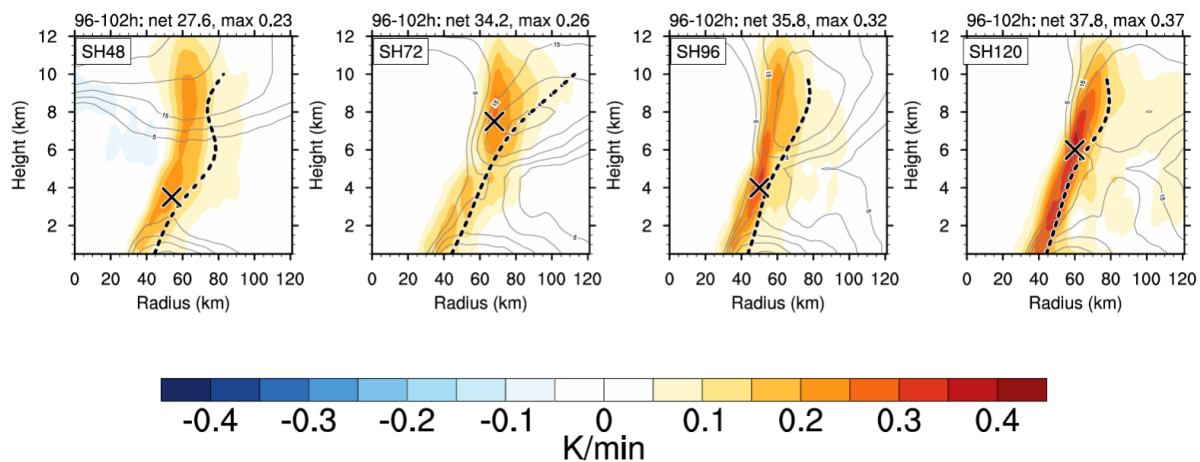


Fig. 17 — As in Fig. 15 (top row), except values are averaged during the period of near-steady intensity 96–102 h after shear starts increasing.

3.2 Cyclone Impacts on Sea Ice in the Arctic during Summer

3.2.1 Climatology of Short-Term Sea Ice Changes during Summer

We begin by examining the climatological change in SIC and total sea-ice area (SIA) during the summer months within the ECB region. We compute SIA by taking the sum across the ECB region of the product of SIC and the area of each grid point. Figure 18 shows the average change in SIA (A-C) and SIC (D-F) from 1-10 days after all days in each month in the 20-year study period of 1999-2018. Error bars represent ± 1 standard deviation of the change in SIA or SIC at each lag day. As expected during the melt season, both SIA and SIC decreases on average from 1-10 days after a given day in June, July, and August. The 10-day reductions in SIA are largest in June and July (Fig. 18A,B), and the 10-day reductions in local SIC are largest for low-concentration sea ice in July (Fig. 18E). Specifically, grid points with SIC between 20-40% in July experience more than 15% SIC reduction after 10 days on average, indicating almost a complete loss of sea ice at these grid points (Fig. 18E). The rate of SIC loss in the 20-40% SIC bins starts to level off after about 5 days in both July and August due to the limited amount of possible SIC loss that can occur within these bins (e.g. the SIC loss for a grid point with 20% SIC cannot exceed 20%). The remainder of this analysis focuses on the 1-5 day changes in sea ice in order to exclude the longer lag times when the rate of SIC loss begins to level off in the low SIC bins.

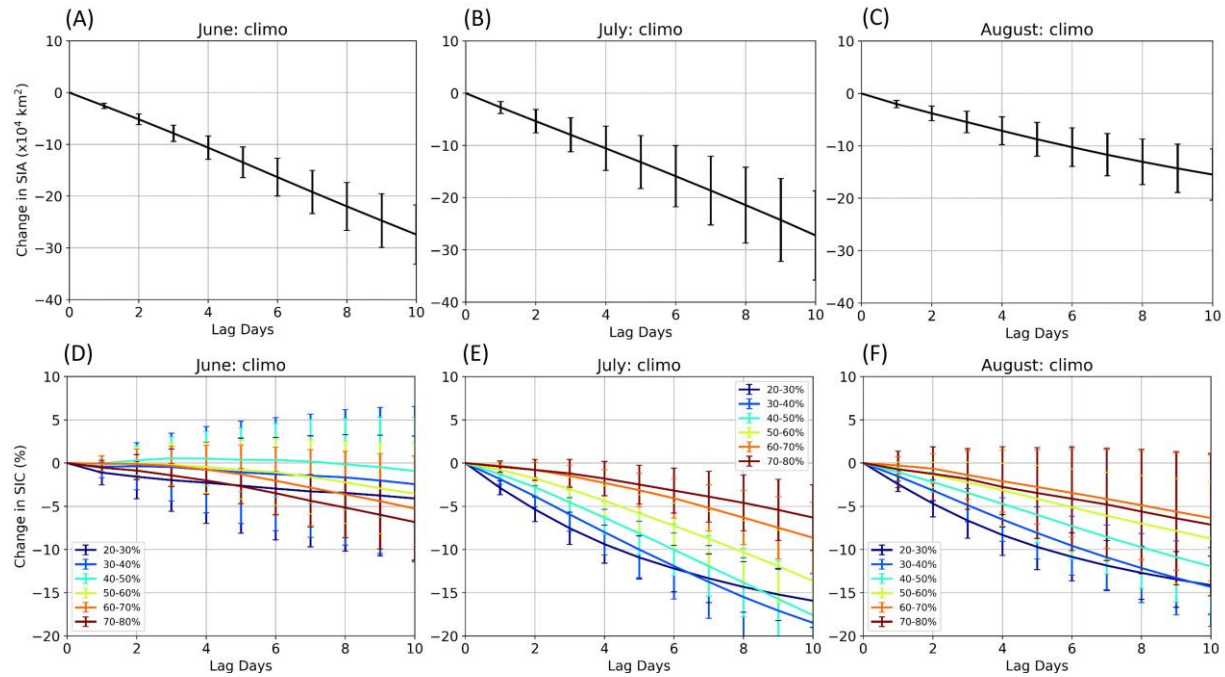


Fig. 18 — Average change in total sea ice area (SIA) within the ECB region (A-C) and the average change in SIC at each grid point in the ECB region and within each 10%-wide bin (D-F) from 1-10 days after a given day in each month. The error bars extend ± 1 standard deviation from the average SIA or SIC change at each lag day.

In order to get a statistical sense for the extent to which each dynamic and thermodynamic forcing is associated with climatological SIC loss in each month, Fig. 19 shows the correlation between 3-day SIC loss in each bin and the SIC advective tendency by the 10-m wind (“adv”), the net shortwave (“SW”) flux at the surface, the net longwave (“LW”) flux at the surface, the surface sensible heat flux (“SH”), the surface latent heat flux (“LH”), and the sum of all four surface fluxes (“net”). Red (blue) colors in Fig. 19 indicate that a forcing term is statistically associated with more (less) local SIC loss (we multiply the advective tendency by -1 in order for this to be the case). In June, stronger SIC advection and, to a lesser

extent, SW flux is associated with enhanced local SIC loss, while stronger LW, SH, and LH fluxes are associated with less SIC loss in June. This is likely due to a strong anticorrelation between the SW flux and each of the LW, SH, and LH fluxes (not shown), and does not actually mean that stronger LW, SH, or LH fluxes cause SIC to increase at a location. In July and August, SIC advection has the strongest relationship to local SIC loss, with larger positive correlations at lower ice concentrations. The relationship between SIC loss and each of the different atmospheric energy fluxes in July and August is similar to June for high concentration sea ice. However, for SIC between 20-40%, stronger LW, SH, and LH fluxes start to be weakly associated with more SIC loss later in the melt season. Nevertheless, the net atmospheric energy flux explains no more than about 2% of the variance in local SIC changes throughout the melt season. Among the forcing terms considered here, the wind-induced SIC advection has the strongest and most consistent relationship to local SIC loss during the summer months.

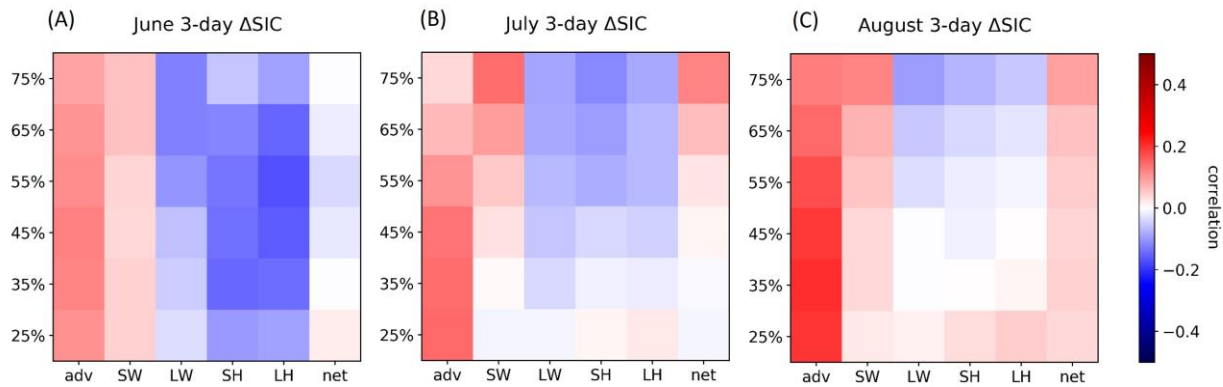


Fig. 19 — Pearson correlation coefficients in June (A), July (B), and August (C) between the 3-day change in sea ice concentration (Δ SIC) and the negative SIC advective tendency (“adv”), the net shortwave flux at the surface (“SW”), the net longwave flux at the surface (“LW”), the sensible surface heat flux (“SH”), the latent surface heat flux (“LH”), and the total surface energy flux from atmosphere (“net”). Correlations are computed separately for grid points in each 10%-wide SIC bin, labeled along the left side of each panel according to the mid-point SIC value in each bin.

If we instead compute correlations between the loss of total SIA within the ECB region and area-averages of each forcing term (Fig. 20), we find somewhat different statistical relationships than for local SIC loss. In June and July, SIC advection has a moderate correlation with 1-day SIA loss and a much weaker correlation with 5-day SIA loss. By August, SIC advection has a stronger correlation to 4-5 day SIA loss than in June and July. This statistical signal of a near-instantaneous response of sea ice cover to wind-induced advection of sea ice early in the melt season, followed by a more lagged response later in the melt season is consistent with the findings of Finocchio et al. (2020). Similar to the SIC correlations shown in Fig. 19, increased SW flux is associated with more SIA loss in June and July. Because the SW flux is the largest component of the surface atmospheric energy balance during the summer months, correlations between SIA loss and the net flux closely reflect those for the SW flux in all three months. In June and July, the magnitude of the correlation between the SW flux and SIA loss increases at longer lag times, indicating that SIA exhibits a lagged response to anomalous surface energy forcing from the atmosphere. This is in contrast to the near-instantaneous response of SIA to the surface wind forcing (i.e. advection), suggesting that surface energy forcing affects sea ice on longer time scales than surface wind forcing during June and July. By August, both the SW flux and the net flux are anti-correlated with SIA loss at all lag times. This negative correlation between SW flux and SIA loss in August suggests that the SW flux is no longer the key component of the surface atmospheric energy budget for understanding short-term SIA changes later in the melt season.

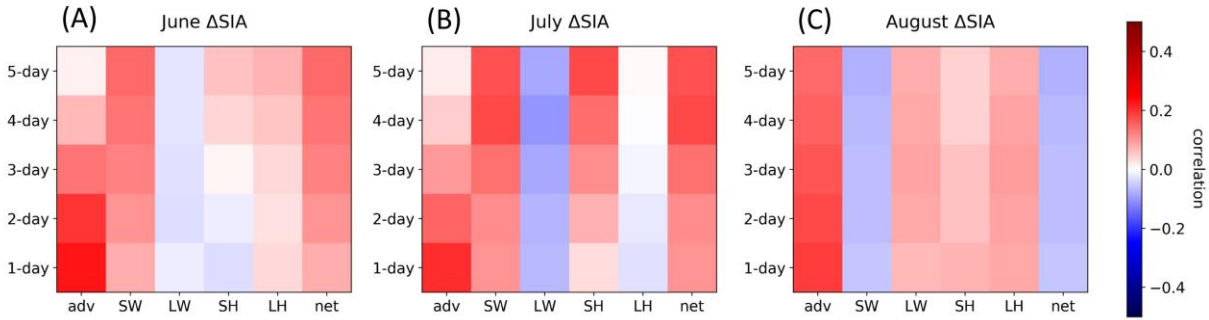


Fig. 20 — Pearson correlation coefficients in June (A), July (B), and August (C) between the 1-5 day change in total sea ice area (ΔSIA) within the ECB region and each forcing term shown in Fig. 19. The y-axis on each panel indicates the lag time over which SIA is computed.

Physically speaking, an increase in any of the surface energy forcing terms should result in enhanced ice loss because an increase in the energy flux from the atmosphere into the sea ice will increase surface melt. Similarly, stronger negative sea ice advection should locally decrease SIC. Therefore, any instances of negative correlations in Figs. 19 and 20 are most likely due to covariances between the forcing terms and do not reflect a causal link to sea ice loss, while positive correlations are more likely to indicate the factors that actually cause ice loss. By this reasoning, the advective tendency and the SW forcing are the most likely forcing terms to cause SIA loss in June and July, while the advective tendency and the combination of LW, SH, and LH fluxes are the most likely forcing terms to cause SIA loss in August. Advection and SW flux (for high concentration sea ice) appear to be the only two terms that consistently have a potential causal link to local SIC loss (Fig. 19).

3.2.2 Statistical Analysis of Summer Cyclone Effects on Sea Ice

The statistical relationships between each forcing term and both local (Fig. 19) and regional (Fig. 20) changes in sea ice provide context for understanding how cyclones affect sea ice, as cyclones locally change the surface energy balance in sea ice-covered areas and induce anomalous wind forcing on sea ice. Figure 21 shows the differences in the average 1-5 day change in SIA (A-C) and SIC (D-F) between cyclone and non-cyclone days in each month during the study period, where non-cyclone days are simply the days in which there is no valid cyclone case inside or within 300 km of the ECB region boundary. The presence of a cyclone is associated with an average of about $3 \times 10^4 \text{ km}^2$ less SIA loss over a 5-day period in June and July compared to days without cyclones. In August, however, the presence of a cyclone is associated with about $1 \times 10^4 \text{ km}^2$ more one-day SIA loss, on average. The smaller amount of SIA lost from 1-5 days after cyclones in June is greater than the climatological variability in SIA changes (represented by the error bars in Fig. 21A extending ± 1 standard deviation from the mean values). Similarly, the increased 1-day SIA loss in August that is associated with cyclones is also outside of the range of climatological variability in SIA changes in August.

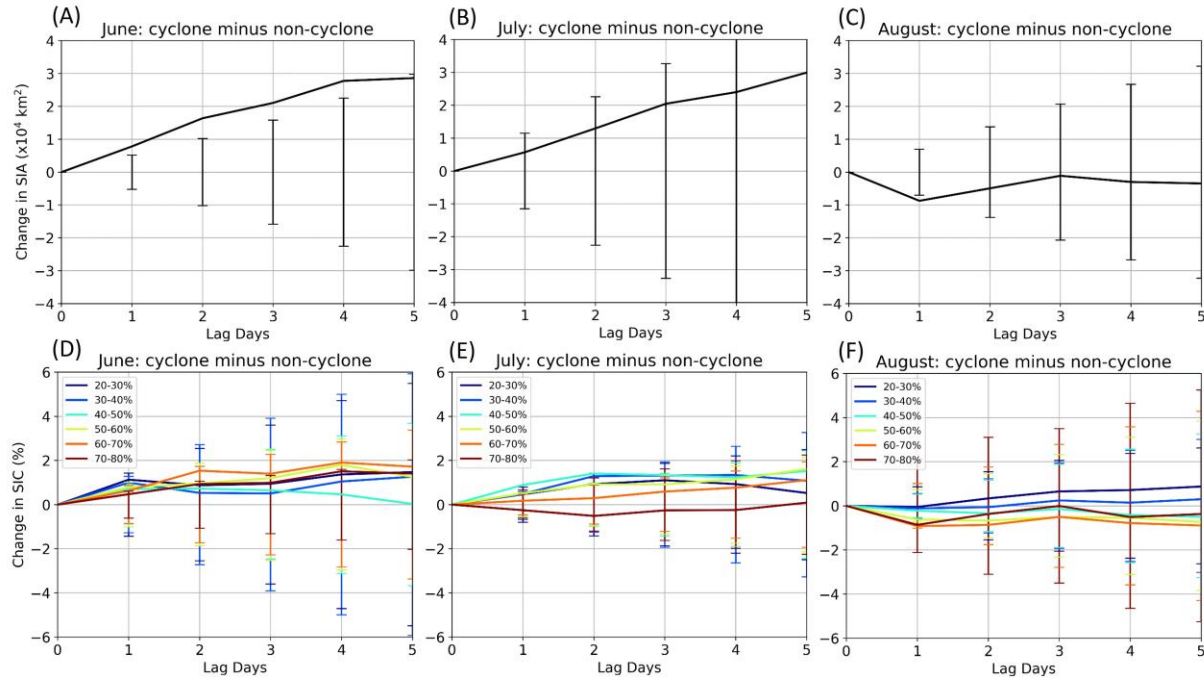


Fig. 21 — (A-C) Difference in the 1-5 day change in SIA in the ECB region between cyclone and non-cyclone days in each month. Error bars are the same as in Fig. 18A-C, and indicate ± 1 -standard deviation in SIA change at each lag time in each month. (D-F) Difference in the 1-5 day change in SIC at each grid point in the ECB region between cyclone and non-cyclone days. Each line indicates difference for grid points with SIC in each 10%-wide SIC bin. Error bars are the same as in Fig. 18D-F, and indicate ± 1 -standard deviation in local SIC change at each lag time in each month. Differences are considered significant if they are outside of the error bars.

The presence of a cyclone in June and July is also associated with less local SIC loss, while cyclones in August are associated with slightly more local SIC loss (Fig. 21D-F). Cyclones decelerate SIC loss by about 1% in the 20-30% SIC bin in June, and by slightly more than 1% in the 20-50% SIC bins in July. Cyclones in July and August, however, are also associated with an accelerated SIC loss by about 1% for higher concentration sea ice (Fig. 19D-F). The differences in local SIC changes between cyclone and non-cyclone days are rarely outside the range climatological variation of SIC changes in each month, indicating that cyclones have a less significant effect on 1-5 day changes in local SIC than they do on the 1-5 day changes in the region-wide SIA.

The key takeaways from Fig. 21 are that (1) cyclone days in June and July have less local and region-wide ice loss compared to days without cyclones, (2) cyclone days in August have slightly more local and region-wide ice loss compared to days without cyclones, and (3) the greater ice loss following August cyclones only persists for about one day after the cyclone day. The remainder of this section focuses on the cyclone days in order to understand the cyclone-related surface forcings that are responsible for the reduced 1-5 day ice loss in June and July and the accelerated 1-day ice loss in August.

Figure 22 shows differences in the average 1-5 day change in SIC between grid points in the upper and lower quartiles of each component of the surface atmospheric energy flux on cyclone days. In June, grid points in the upper SW flux quartile experience significantly more 1-5 day SIC loss than grid points in the lower SW flux quartile, regardless of the initial SIC (Fig. 22A). In contrast, grid points in the upper quartiles of both LW flux (Fig. 22D) and the surface heat flux (Fig. 22G) experience significantly *less* SIC loss than grid points in the lower quartiles of each flux. This unintuitive relationship in which larger surface heat fluxes and LW fluxes are associated with decreased SIC loss suggests that these fluxes do not drive the local changes in SIC, but co-vary with other surface forcings that do. For example, the cloud radiative forcing at the surface is negative in the shortwave and positive in the longwave. Because the negative SW anomalies due to clouds in June tend to be larger in magnitude than the positive LW

anomalies (e.g. Finocchio et al. 2020), cloudy grid points frequently end up having less ice loss despite having positive LW flux anomalies.

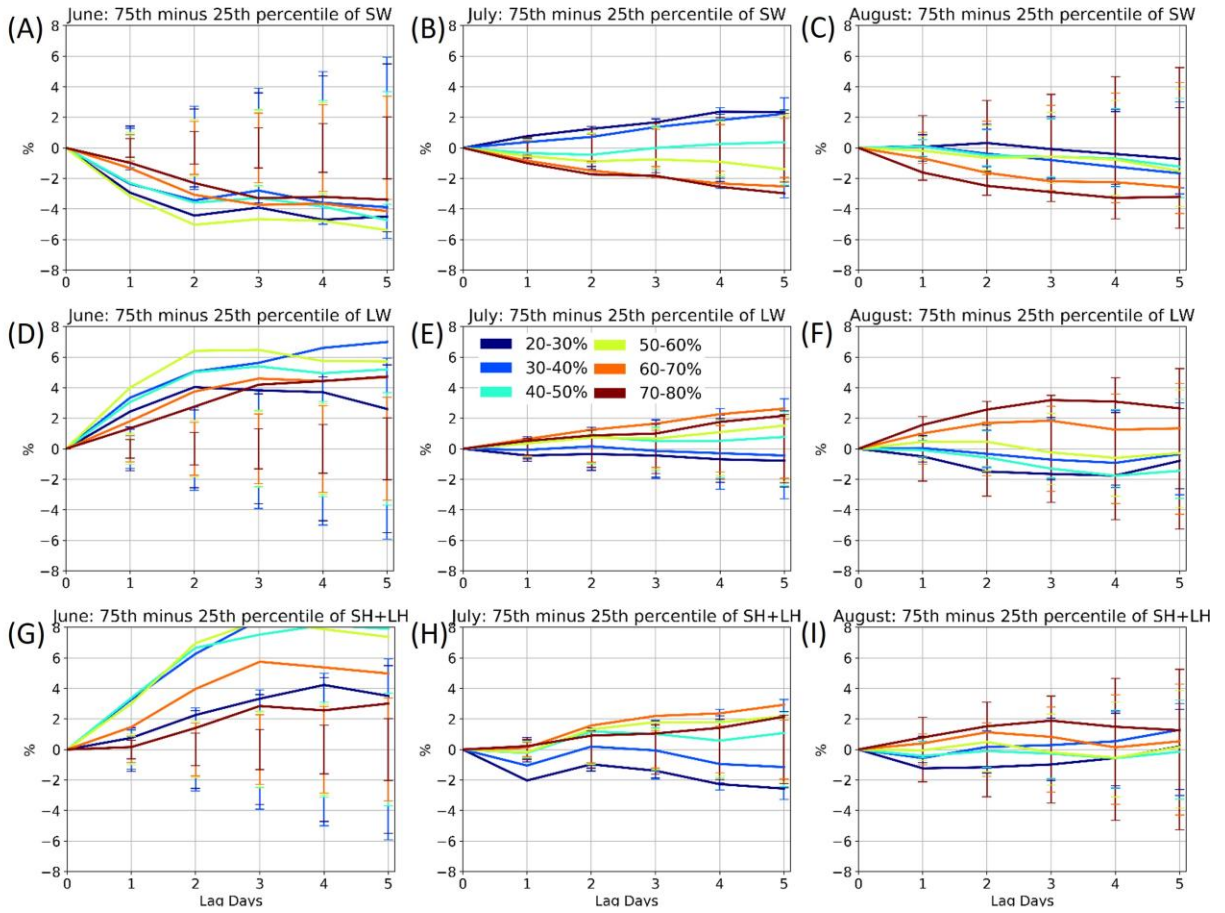


Fig. 22 — Differences in each month in the 1-5 day change in SIC concentration averaged over grid points in the upper versus lower quartiles of the SW flux (A-C), LW flux (D-F), and surface heat flux (H-J). As in Figs. 18 and 21, line colors indicate the different 10%-wide SIC bins. Error bars are the ± 1 standard deviation of each flux term for grid points with SIC in each bin computed over all days in each month for the 20-year study period.

In July and August, the difference in SIC change between the upper and lower quartiles of SW flux depends on the SIC; strong SW fluxes are associated with less 1-5 day SIC loss for initially low-concentration sea ice, and more SIC loss for initially high-concentration sea ice (Fig. 22B,C). This suggests that as the melt season progresses, stronger SW fluxes continue to be an important factor in accelerating the local SIC loss for initially high ice concentrations, while other factors begin to play a greater role in accelerating local SIC loss at lower ice concentrations. One of those factors is the surface heat flux. Low ice concentration ($\text{SIC} < 40\%$) grid points in the upper quartile of surface heat flux in July and August experience 1-2% more 1-day SIC loss on average than grid points in the lower quartile of surface heat flux (Fig. 22H,I). The surface heat flux is a function of both the surface wind speed and the surface temperature and moisture disequilibrium. We find that the 2-m air temperature differentiates grid points in the upper versus lower quartiles of the surface heat flux better than the 10-m wind speed; grid points with ice concentration $< 40\%$ that are in the upper heat flux quartile in July and August are consistently 2°C warmer, on average, than grid points in the lower heat flux quartile, while the average 10-m wind speed of grid points in each quartile differs by less than 1 m s^{-1} . This indicates that the warm air masses that Arctic cyclones transport poleward into the Arctic (e.g. Fearon et al. 2020) are capable of

accelerating the short-term loss of low-concentration sea ice later in the melt season by locally increasing the surface heat flux over sea ice.\

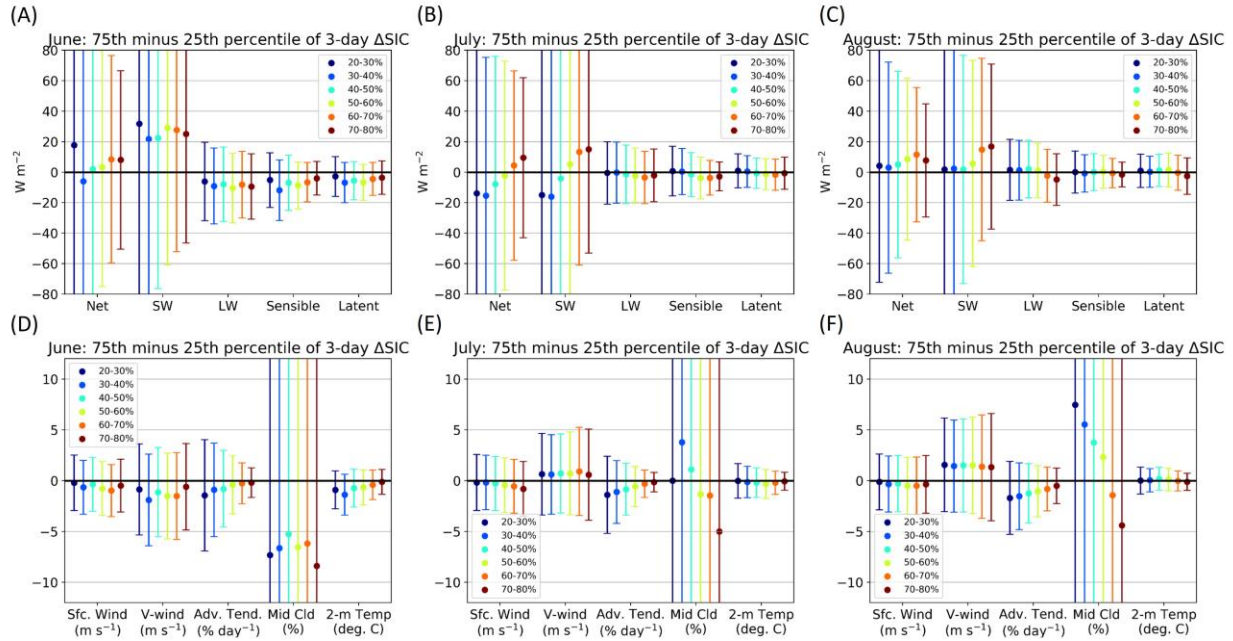


Fig. 23 — Differences in each term in the surface energy budget (A-C) and in the surface wind, advective tendency, mid-level cloud, and 2-m temperature (D-F) between grid points in the upper and lower quartiles of 3-day SIC loss. Differences are computed separately for grid points within each 10%-wide SIC bin. Error bars indicate ± 1 standard deviation in each term computed over all days in each month for the 20-year study period.

Figure 23A-C shows the differences in each term in the surface energy budget between grid points in the upper versus lower-quartiles of 3-day SIC change on cyclone days, where the upper quartile here represents the largest local SIC loss. We note that in all three months, differences in each flux are never larger than the climatological variability of each term (indicated by error bars extending ± 1 standard deviation from the mean values), which indicates that factors other than the local surface energy flux from the atmosphere also contribute to local SIC variability during the summer. Nevertheless, these differences are still useful for understanding the relative contributions of each atmospheric flux to local SIC loss. In June, the net flux is $\sim 19 \text{ W m}^{-2}$ larger on average for grid points in the upper quartile of SIC loss in the 20-30% SIC bins. This larger net flux is due to a larger SW flux that is only partially offset by a smaller LW and sensible heat flux. For the grid points with SIC between 30-60%, however, the larger SW fluxes are completely offset by reduced LW and surface heat fluxes. In July and August, differences in the average LW and surface heat fluxes between grid points in the upper and lower quartiles of SIC loss are generally less than 5 W m^{-2} , and differences in the net flux are due primarily to differences in the SW flux. Low-concentration sea ice grid points that experience the most SIC loss have net and SW fluxes that are $10\text{-}15 \text{ W m}^{-2}$ smaller than grid points that experience the least SIC loss. For higher concentration sea ice in July and August, however, the largest SIC loss is associated with $10\text{-}15 \text{ W m}^{-2}$ larger SW fluxes. This is consistent with Figs. 22B and 22C, which showed that larger SW fluxes in July and August are only associated with enhanced sea ice loss for high concentration sea ice.

Figure 23D-F shows differences in the 10-m wind speed, the 10-m meridional wind, the mid-level (800-450 hPa) cloud fraction, and the 2-m air temperature between grid points in the upper versus lower quartiles of SIC loss on cyclone days. We only consider mid-level clouds because of their relatively large SW radiative forcing and because their variability is more closely tied to cyclones than the fog and low-level stratiform clouds that are prevalent in the Arctic during summer (Eastman and Warren 2010).

Similar to the surface fluxes, differences in each of the variables shown in Fig. 23D-F are never larger than the climatological variability, but are still useful for understanding the relative contributions of each variable to anomalous local SIC loss. In June, surface winds tend to be slightly weaker and less southerly on average at grid points with the most SIC loss. The SIC advection by the 10-m wind is more negative by about 1.5% per day for low concentration sea ice. Furthermore, the 2-m temperature is cooler by up to 1.5 C and the mid-level cloud fraction is 5-8% lower at grid points with the most SIC loss, which is consistent with a larger SW flux (Fig. 23A). Together, these results suggest that the largest 3-day reductions in local SIC during June tend to occur where conditions are calmer, clearer, and cooler. By July and August, the surface wind speed and surface temperature are strikingly similar between grid points with the largest and smallest 3-day SIC loss, but the meridional wind component tends to be more southerly for the largest SIC loss. The differences in the mid-level cloud fraction have more of a dependence on the SIC than in June. For high-concentration sea ice, large local SIC loss is associated with reduced mid-level cloud fraction (Fig. 23E,F) and increased SW flux (Fig. 23B,C), as in June. However, the largest local reductions in low-concentration sea ice in July and August are associated with a larger cloud fraction and smaller SW fluxes. This combined with the increased southerly wind and more negative SIC advection by about 2% per day on average for low-concentration sea ice in July and August suggests that cloudy, southerly flow regimes increasingly accompany large local SIC loss events in the Arctic as summer progresses.

Certain surface wind regimes may promote the strong, negative SIC advection that is frequently associated with large reductions in low-concentration sea ice. To further explore this, we compute differences in the total wind speed and each component of the local wind vector between cases in the upper versus lower quartiles of SIC advection (Fig. 24). In all three months, grid points that experience strong SIC advective tendencies have an average total and zonal wind speed that is similar to grid points that experience weak SIC advection. However, the average meridional wind speed is up to 1.5 m s^{-1} larger (i.e. more southerly) at grid points with the largest negative SIC advection. The more southerly wind component associated with strong negative SIC advection also becomes more apparent as the melt season progresses. Thus, the meridional wind component is what determines how much negative SIC advection occurs within the ECB region – particularly later in the melt season when low concentration sea ice is prevalent. The preference for large negative SIC advection under southerly wind conditions likely relates to the primarily northward-directed SIC gradient within this region of the Arctic (Fig. 4).

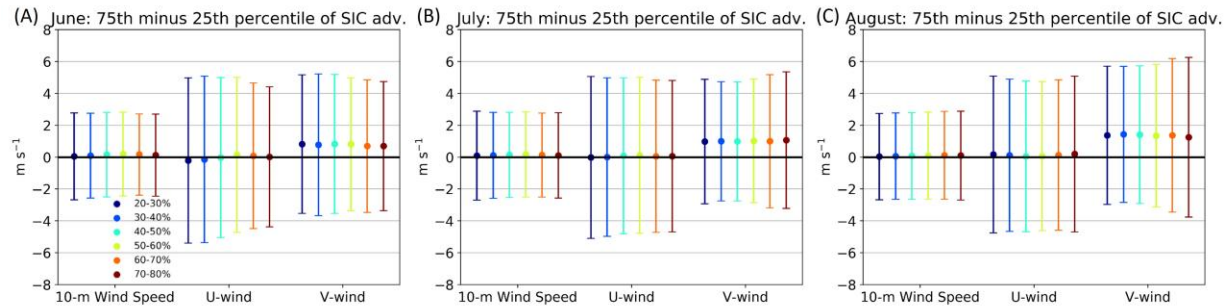


Fig. 24 — Differences in the 10-m wind speed and the u- and v-wind components between grid points in the upper and lower quartiles of SIC advective tendency in each SIC bin. Error bars indicate ± 1 standard deviation in each term computed over all days in each month for the 20-year study period.

3.2.3 Comparison of an Early and a Late Season Cyclone

A key result from the previous section is that the surface forcings that differentiate large from small local and regional ice loss events vary over the course of the melt season and depend to some extent on the initial ice concentration. In this section, we further explore the seasonal dependence of cyclone impacts on sea ice by comparing two cyclone cases that impacted the same region of the Arctic early and

late in the melt season. As in the previous section, the analysis of cyclone impacts focuses on the surface wind and energy forcing.

The two selected cyclones occurred in June 2012 and August 2016. Both cyclones reached a similar maximum intensity and affected sea ice in the East Siberian Sea. The June 2012 cyclone emerged from Siberia on 19 June and intensified as it moved toward the northeast, reaching a maximum intensity of 970 hPa at 12Z on 20 June. The August 2016 cyclone was located further north and translated eastward from the Laptev Sea, reaching a maximum intensity of 968 hPa at around 12Z on 15 August and maintaining that intensity for several hours as it made a loop into the central Arctic. Figure 25A,B shows the cyclone tracks and the sea level pressure contours at 12Z on the day when the cyclones were near their peak intensity. The color shading denotes the net surface energy flux anomalies relative to the 1999-2018 average. Both cyclones have areas where the net flux over sea ice is more than 20 W m^{-2} higher than the 20-year average. In both cases, the positive net flux anomalies are primarily associated with positive anomalies in the net longwave, sensible, and latent heat fluxes, while the negative net flux anomalies are associated with negative anomalies in the net SW flux (not shown). The middle panels (25C,D) show SIC and daily averaged 10-m wind vectors, and reveal that positive surface flux anomalies tend to coincide with where the surface winds are directed off of the Eurasian continent. The bottom panels (25E,F) show the distributions of SIC within the boxes outlined in Fig. 25A-D. The June cyclone primarily affected sea ice with concentrations $> 60\%$, much of which was adjacent to the Siberian coast. Consequently, internal ice forces were likely to be considerably stronger in this case than in August, greatly reducing the wind-induced drift of sea ice. In contrast, the August 2016 cyclone affected lower-concentration sea ice that was further from the Eurasian coastline. This made the sea ice more susceptible to drifting in response to the surface winds of the cyclone.

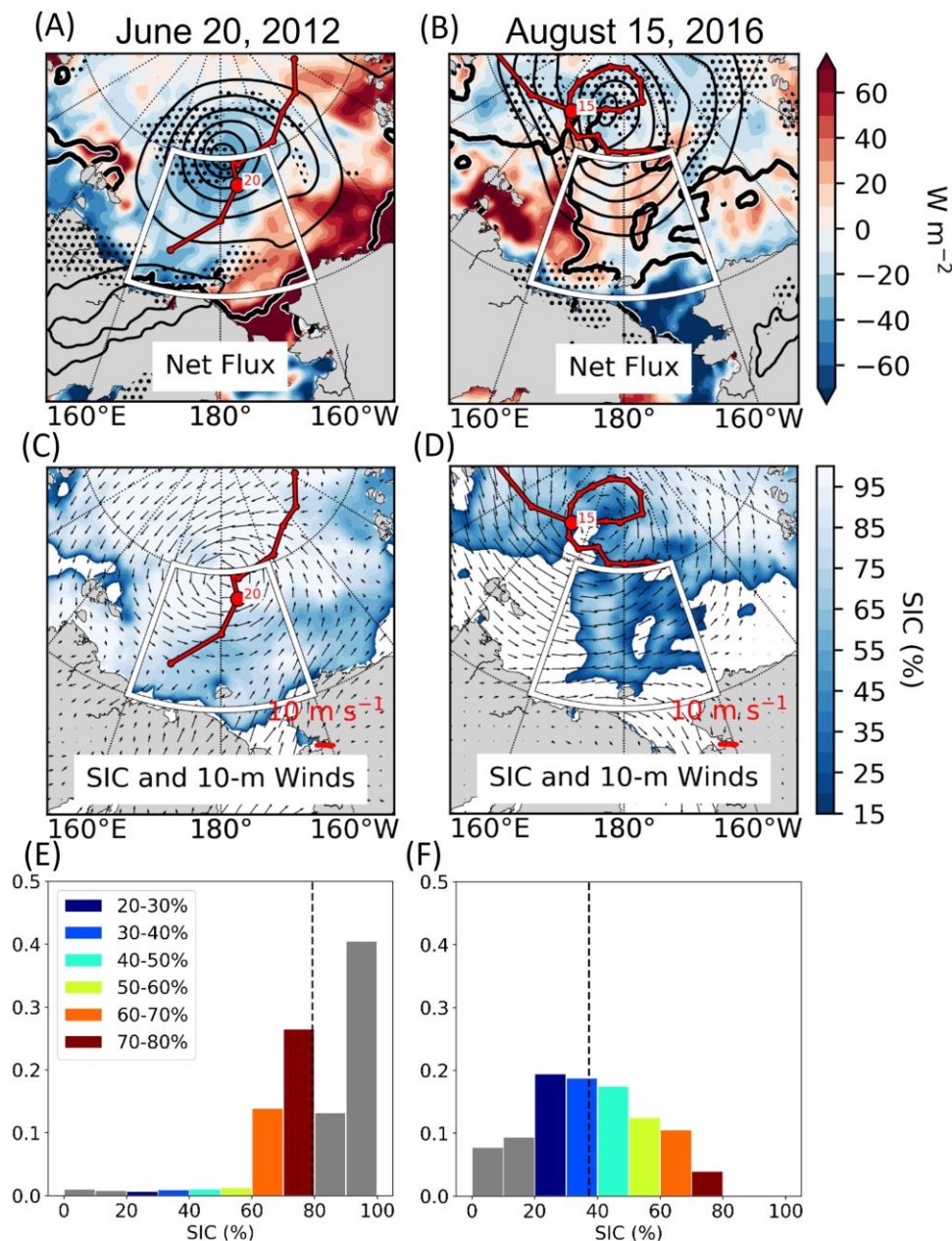


Fig. 25 — (A,B) Cyclone tracks (red line, 00Z cyclone position labeled by day of month), daily averaged net surface energy flux anomalies (shading) and mid-level cloud fraction > 70% (black stippling), and sea level pressure at 12Z on 20 June 2012 (A), and 15 August 2016 (B). The thick black line is the 15% SIC contour indicating the sea ice edge. Sea level pressure contours are drawn every 5 hPa less than 1000 hPa. (C,D) Sea ice concentration (shading), cyclone track (red line) and daily averaged surface wind vectors on 20 June 2012 (C), and 15 August 2016 (D). (E,F) Probability density of SIC within the white outlined region in A-D on each date colored according to SIC bins as in Fig. 4. The vertical dashed line in E,F denotes the average SIC within the outlined region.

Figure 26A,B shows times series of the area-weighted average within the outlined regions in Fig. 25 of the surface energy flux and the mid-level cloud fraction anomalies (relative to the 20-year average) for each cyclone. Quantities are only averaged over grid points covered by sea ice or open water (i.e. land points are excluded). Figure 26C,D shows time series of the maximum 10-m wind speed, 2-m temperature, and SST averaged in the same region as in Fig. 26A,B. During both cyclones, the largest

positive anomaly in the net surface energy flux occurs about 6 h after the largest positive anomaly in mid-level cloud fraction. During the June cyclone, surface flux anomalies reach a local maximum about 12 h before the peak 10-m wind speeds occur, while the peak wind speed and surface flux anomalies roughly coincide during the August cyclone. In both cyclones, the sensible heat flux is the largest positive anomalous energy forcing, which is likely due to the combination of strong surface winds and above average 2-m temperature. The largest negative anomalies in the net SW flux associated with the cloud cover from each cyclone case occur at local noon (~00Z) on 20 June and 15 August, respectively. Despite clouds reducing the downward SW flux by a larger amount in June, the magnitude of the negative anomalies in the net SW flux is similar during the June and August cyclones, which is likely because the area-averaged surface albedo in June (0.55 on 20 June) is so much larger than in August (0.18 on 15 August).

Figure 26E,F show times series of the SIC distributions. In August, the initial SIC distribution has a much higher frequency of low-concentration sea ice compared to in June. Furthermore, the sea ice area in August 2016 (thick black line in Fig. 26F) was already 30-50% lower than the 20-year average ice area within this sub-region (dotted black line). As a result, both the reductions in sea ice area and the shift toward lower concentration sea ice associated with the August 2016 cyclone are more apparent than for the June 2012 cyclone. The June 2012 cyclone also induces a distributional shift toward lower concentration sea ice, but has almost no effect on the total sea ice area (Fig. 26E). Moreover, the distributional shift in SIC during the June cyclone is only temporary, as the SIC distribution begins to resemble the pre-cyclone distribution two days after the passage of the cyclone. These differences in the apparent sea ice impacts of two cyclones that have similar intensity and similar effects on the surface energy balance emphasize the importance of the initial sea ice state for anticipating how cyclones will affect sea ice on short time scales (Lukovich et al. 2021).

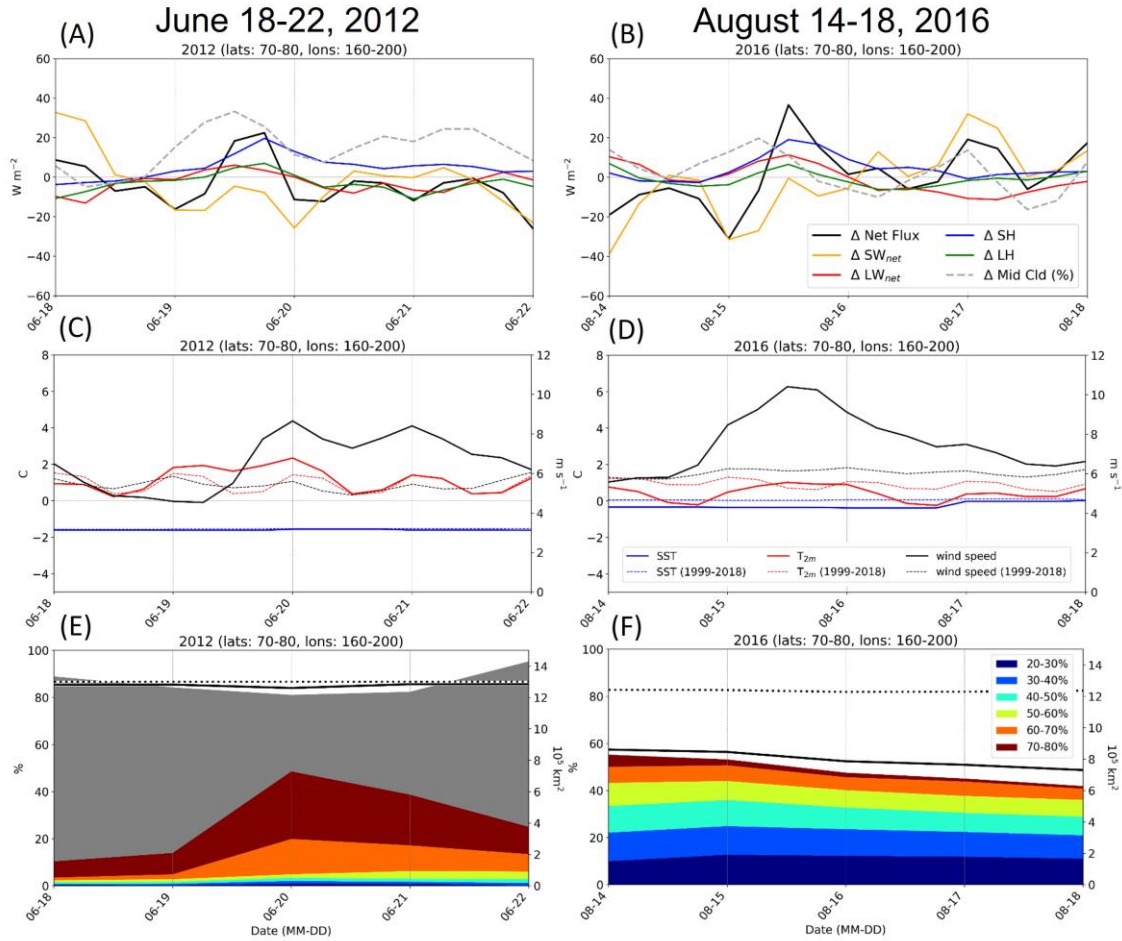


Fig. 26 — (A,B) time series of the area-weighted average of anomalies in each term in the surface energy budget ($W m^{-2}$) and the mid-level cloud fraction (%) from 18-22 June 2012 (A) and 14-18 August 2016 (B). Averages are taken within the region outlined in white in Fig. 25A-D, and anomalies are computed with the respect to the 20-year average of each quantity in this region. (C,D) As in A,B, but for the average SST and 2-m air temperature (left y-axis), and 10-m wind speed (right y-axis) for June 2012 (C) and August 2016 (D). Solid lines denote averages computed in the year in which the cyclone occurred and dotted lines denote the 20-year averages. (E,F) The total fraction of grid points that are in each SIC bin during the same period in June 2012 (E), and August 2016 (F). Color-filled areas indicate the fraction of the region covered by grid points with SIC between 20% and 80%. Gray (white) regions indicate the fraction of area covered by SIC > 80% (SIC < 20%). The solid black line is the total sea ice area and the dotted black line is the 20-year average sea ice area in the region (right y-axis).

Figure 27 shows scatter plots of the 5-day change in SIC versus the various surface forcing terms computed from ERA-5 within the boxed region outlined in Fig. 25A,B, and for grid points with SIC between 20% and 80%. The surface energy fluxes in Fig. 27A-F and the winds used to compute the advective tendencies in Fig. 27G,H are averages of ERA-5 fields at four 6-hourly time steps in each day. In order to increase the sample size, the scatter plots include data from two days for each cyclone case during which there were positive net surface flux anomalies in Fig. 26A,B (19 and 20 June for the 2012 cyclone, and 15 and 16 August for the 2016 cyclone). Points in Fig. 27 that correspond to grid points in the upper decile of total cloud fraction are outlined in black.

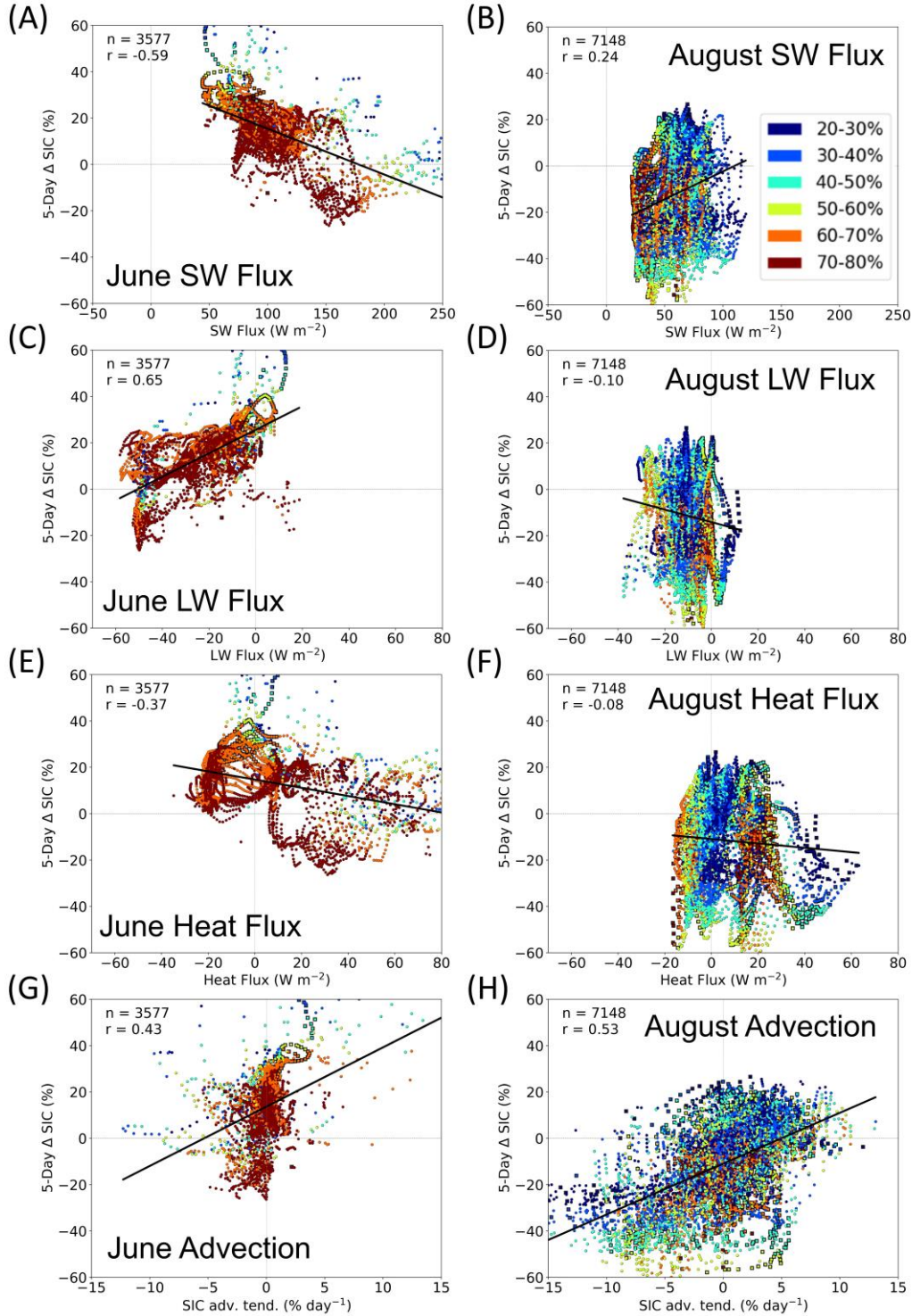


Fig. 27 — Scatter plots of 5-day change in SIC at each grid point and the net SW flux (A,B), the net LW flux (C,D), the surface sensible + latent heat flux (E,F), and the SIC advective tendency (G,H) on 19-20 June, 2012 (left column) and on 15-16 August 2016 (right column). Grid points are colored according to the SIC bin. Only grid points with SIC between 20% and 80% are plotted. Grid points that are in the upper decile of total cloud fraction are outlined in black. The black line is the linear fit through the data. The number of grid points (n) and the linear correlation (r) is printed in the upper left corner of each plot. Note the different x-axis ranges in each row.

The relationship between the SW flux and changes in SIC varies considerably between the June and August cyclones. For the June cyclone, the relationship between local SIC change and SW flux (Fig. 27A) is negative ($r=-0.59$), indicating that stronger SW flux is associated with enhanced local SIC loss. The sign of the relationship is consistent with climatology (Fig. 19A), but the magnitude is much larger. In contrast, the relationship between local SIC change and SW flux for the August cyclone is weakly positive ($r=0.24$, Fig. 27B). Tables 3 and 4 list the average values of each flux over all the grid points shown in Fig. 27 (“All”), grid points with 5-day SIC loss (“Ice Loss”), and grid points in either the lower decile of total cloud fraction (“Clear”) or the upper decile (“Cloudy”). Only a small fraction of the grid points in June experience SIC loss, but among those, the average SW flux is 41% larger than the average SW flux over all grid points in Fig. 27A (Table 3). In contrast, the SW flux among the much larger number of grid points that experience SIC loss in August (60 W m^{-2}) is similar to the overall average (63 W m^{-2}), suggesting that the SW flux plays less of a role in local SIC changes in August. The smaller role of the SW flux in modulating sea ice variability in August may relate to a weaker cloud SW radiative forcing during the August cyclone. We estimate the cloud SW radiative forcing by comparing the downward component of the SW flux between the clear and cloudy grid points for each cyclone case. In June, the average downward SW flux is 371 W m^{-2} for clear grid points and 135 W m^{-2} for cloudy grid points, so clouds reduce the downward SW flux at the surface by 64%. During the August cyclone, clouds only reduce the downward SW flux by 38% relative to clear skies. Thus, the cloud SW radiative forcing is 26% smaller for the August cyclone than for the June cyclone. Because the average surface albedo is similar for clear and cloudy grid points in June and August, the reduced cloud SW radiative forcing in August means that clouds associated with the August cyclone have a significantly smaller impact on how much SW radiation is absorbed at the surface than clouds associated with the June cyclone.

In terms of the LW flux, we find a strong positive relationship between SIC change and LW flux in June (Fig. 27C) and a weak negative relationship in August (Fig. 27D). The positive relationship in June, in which stronger (or less upward) LW fluxes are associated with less SIC loss, is consistent with the climatology in this region (Figs. 19A and 20A) and, as discussed earlier, is likely a result of strong covariances between the LW and SW flux. The average LW flux in clear conditions is more negative than in cloudy conditions by 43 W m^{-2} in June (Table 3), which on its own would favor less ice loss, but the 105 W m^{-2} increase in SW flux in clear conditions compared to cloudy conditions more than compensates for the reduced LW flux. In August, the averages of both the SW and LW fluxes over ice loss grid points are similar to the respective averages over all grid points. In addition, the net radiative flux (SW+LW) in cloudy conditions is only about 13 W m^{-2} larger than in clear conditions in August (Table 4), providing further support for the idea that clouds associated with the June cyclone are much more likely to affect sea ice by changing the surface energy balance than clouds associated with the August cyclone.

Table 3 — Average values of each surface energy flux and SIC advective tendency over different sets of grid points shown in Fig. 27A,C,E,G during the June 2012 cyclone: All grid points (“All”), only grid points with 5-day SIC loss (“Ice Loss”), grid points in the upper decile of total cloud fraction (“Cloudy”) and grid points in the lower decile of total cloud fraction (“Clear”).

June 2012	SW (W m^{-2})	LW (W m^{-2})	Heat (W m^{-2})	Net (W m^{-2})	SIC Adv. ($\% \text{ day}^{-1}$)
All	111	-24	8	95	-0.2
Ice Loss	157	-40	34	150	-1.1
Cloudy	73	-5	-3	65	0.5
Clear	178	-48	31	161	-1.6

Table 4 — As in Table 3, but for the grid points shown in Fig. 27B,D,F,H during the August 2016 cyclone.

August 2016	SW (W m ⁻²)	LW (W m ⁻²)	Heat (W m ⁻²)	Net (W m ⁻²)	SIC Adv. (% day ⁻¹)
All	63	-10	4	57	-0.2
Ice Loss	60	-10	4	54	-1.3
Cloudy	52	-5	9	56	0.5
Clear	82	-22	-4	56	0.4

Although the surface sensible and latent heat fluxes contribute substantially to the positive surface energy flux anomalies associated with the cyclones (Fig. 26A,B), the surface heat flux on its own has only a weak relationship to local SIC loss in both the June and August cyclones (Fig. 27E,F). This is because the magnitude of the heat flux is generally much smaller than that of the net radiative flux (Tables 3 and 4). Nevertheless, the average heat flux for ice loss grid points in June is 26 W m⁻² larger than the average over all grid points, indicating a supporting role of stronger heat fluxes in local SIC loss. The average heat flux is larger at clear grid points for the June cyclone (Table 3), and at cloudy grid points for the August cyclone (Table 4). However, the increased heat flux in cloudy conditions during the August cyclone does not appear to be the main cause of local SIC loss within the region of interest, as the average heat flux for ice loss grid points during the August cyclone is the same as the average over all grid points (Table 4).

Figures 27G and 27H show the relationship between 5-day SIC change and the SIC advective tendency from the 10-m winds. For both the June and August cyclones, the sign of the relationship is positive and significant at the 98% confidence level. In addition, higher concentration sea ice tends to be associated with the weakest advective tendencies, while lower concentration sea ice is associated with stronger advective tendencies. Because the June cyclone primarily impacts higher concentration sea ice, the positive relationship between SIC advection and the local change in SIC is somewhat weaker than that for the August cyclone ($r=0.43$ in June vs. 0.53 in August). This is consistent with Fig. 19, which shows that the magnitude of the correlation between SIC loss and the negative SIC advection is largest for low concentration sea ice in August. Nevertheless, the average advective tendency over ice loss grid points for both of the cyclone cases is more negative than the average over all grid points by about 1% per day (Tables 3 and 4). Unlike the surface energy fluxes, whose contributions to local SIC loss vary considerably between the June and August cyclones, sea ice advection has consistent relationship to the local changes in SIC for both cyclone cases.

4. CONCLUSIONS

4.1 The Effect of Increasing Vertical Wind Shear on Tropical Cyclone Rapid Intensification

The first part of the research that was part of this Karles Fellowship involved conducting eight simulations in which westerly vertical wind shear increases from 3 to 15 m s⁻¹ at different stages in the lifecycle of an intensifying TC. Increasing shear while the TC is a weak tropical storm (SH24) causes the vortex to vertically decouple and inhibits intensification. For all of the stronger initial TCs, increasing shear causes an initial period of absolute weakening followed by a period during which the TC maintains a near-steady intensity. The amount of initial weakening that occurs in response to increasing shear varies among the simulations. TCs exposed to shear toward the end of RI tend to weaken by more than the TCs exposed to shear earlier in their intensification. The most dramatic weakening occurs in the TC in SH96 exposed to increasing shear at the end of RI, when it is a Category 4 hurricane. Although the TC in SH144 has a similar intensity as the TC in SH96 when the shear increases, it weakens by a smaller amount partly because it has a broader wind field, which supports stronger vertical coupling throughout the lower troposphere. We speculate that the enhanced vulnerability of a TC near the end of RI to increasing shear is a result of the TC nearing its maximum potential intensity while still having a fairly compact tangential wind field (e.g., Fig. 6). Whether this is the case for real TCs or only applies to TCs in these idealized simulations requires further investigation. The near-steady intensity of TCs after adjusting to 15-m s⁻¹ shear also depends on the intensity and size of the TC when the shear increases; initially stronger and larger TCs generally have a higher near-steady intensity long after shear increases than initially weaker and smaller TCs.

The structural response to increasing shear involves an increase in vortex tilt and the development of a wavenumber-1 precipitation asymmetry. Increasing shear causes the initially smaller and weaker TCs to develop larger tilt magnitudes (20–30 km) than the stronger and larger TCs (≤ 10 km). The Rossby penetration depth quantifies vertical coupling of a dry vortex as a function of vortex size, intensity, and inner-core static stability. Despite limitations in applying such a metric to the moist convective vortices analyzed in this study, we find it still provides useful insight into the different tilt responses among the simulations. Increasing shear when the TC has a large penetration depth later in its lifecycle causes the vortex to tilt considerably less than increasing shear at an earlier developmental stage. The increase in the Rossby penetration depth is largely due to the increasing intensity of the TCs and, to a lesser extent, the increasing size of the low- and mid-level vortex. The average static stability of the inner core does not appear to affect vortex tilt as it is similar in all simulations.

Lateral ventilation is found to be a key factor in the response to increasing vertical wind shear. Backward trajectories seeded within different layers of the inner core shortly after the shear increases reveal that the wind profile used in this study promotes the strongest lateral ventilation just below the outflow layer from 8–11 km. Lateral ventilation in this layer initially increases as the initial TC rapidly intensifies and the radial gradients in θ_e outside the inner core become stronger. As a result, the initial weakening of even the strongest TCs in these simulations is likely due in part to lateral transport in the 8–11 km layer that is capable of reducing inner-core θ_e by nearly 2 K. However, further deepening and expansion of the tangential winds of the initial vortex in the SH84–SH144 experiments eventually results in a decrease in mid-level lateral ventilation, consistent with both kinematic models (Riemer and Montgomery 2011) and observations of intense TCs that deflect storm-relative flow around their inner core (e.g., Willoughby et al. 1984, Marks et al. 1992). Lateral ventilation is less active when TCs reach a near-steady intensity at the end of the simulations, but remains capable of reducing inner-core θ_e in the 5–8 km and 8–11 km layers by about 1 K for the TCs with the weakest near-steady intensity, and by lesser amounts for those with stronger near-steady intensities. These inner-core θ_e reductions are upper bounds on the extent to which lateral ventilation destructively affects the simulated TCs because the calculations include air parcels that approach the inner core on elliptical paths around the storm center but do not necessarily get entrained in convective updrafts.

The extent to which increasing shear induces downdrafts that flush low- θ_e air into the boundary layer also depends on the initial intensity of the TC, but in a way that is perhaps not intuitive. Intense TCs nearing the end of RI tend to produce the strongest downward fluxes of low- θ_e air into the boundary layer in response to increasing shear. Consequently, these TCs also have lower θ_e within the USL quadrant of the boundary layer. Further analysis of backward trajectories reveals that, despite a substantial increase in θ_e in the USR quadrant via surface enthalpy fluxes, the average θ_e of downdraft air parcels by the time they ascend from the boundary layer in the DSR quadrant is lower in the intense TCs that experience the strongest flushing (e.g., SH96) than in the TCs with weaker flushing. Thus, the reduced θ_e of air parcels ascending downshear likely contributes to the substantial weakening that occurs in response to increasing shear. During the period of near-steady intensity, however, the average θ_e of air parcels entering downshear updrafts is similar in each simulation, implying that boundary layer thermodynamics do not influence the near-steady intensity of TCs as much as the initial weakening response to shear.

Increasing shear also shifts the strongest diabatic heating outside of the RMW, where it is less efficient at spinning up the tangential winds. The amount by which diabatic heating inside the RMW decreases in response to shear is well correlated ($r > 0.9$) with the amount of weakening that occurs several hours later. At the end of the simulations, the diabatic heating inside the RMW is also well correlated with the near-steady intensity of the TCs, but it is difficult to tell whether the different magnitudes of diabatic heating inside the RMW cause or simply reflect the different near-steady intensities. Notably, the changes in diabatic heating inside the RMW correspond to changes in vortex tilt, demonstrating how increasingly misaligned vortices yield heating configurations that are less optimal for intensification (e.g., Schecter and Menelaou 2020). Diabatic heating, vortex tilt, boundary layer flushing and lateral ventilation likely all cooperate to some degree in order to produce the TC intensity- and size-dependent responses to increasing shear documented in this study.

That TCs retain some memory of their intensity prior to encountering shear raises intriguing questions about the equilibrium intensity of sheared TCs in nature. Tang and Emanuel (2012) subjected axisymmetric TCs with different intensities to a fixed amount of mid-level ventilation and found that all except the weakest TCs eventually reached the same intensity. We suspect that the different near-steady intensities of the TCs in our simulations is a result of prescribing vertical wind shear rather than ventilation, such that the structure and intensity of the TC determines how much a given magnitude of vertical wind shear ventilates the storm. An unrealistic aspect of the simulations in this study is that vertical wind shear increases while all other aspects of the environment remain constant. In reality, covariances between vertical wind shear and humidity or sea surface temperature would prevent most TCs from reaching any sort of equilibrium with their environment (e.g. Finocchio and Majumdar 2017a, Rios-Berrios and Torn 2017). Nevertheless, the results of this study caution against thinking of the destructive effects of environmental factors like vertical wind shear on TCs in terms of universal thresholds, as such thresholds are likely to depend on characteristics of the TC.

4.2 Cyclone Impacts on Sea Ice in the Arctic during Summer

The second part of the research conducted as part of this Karles Fellowship quantifies the effect of cyclone winds and surface energy forcing on sea ice within the Pacific sector of the Arctic during the melt season months of June, July, and August. We consider the climatology of sea ice change during these months, as well as the statistics of cyclone effects on sea ice for a large sample of cyclone cases and for two individual cyclone cases that impacted sea ice in the East Siberian Sea near the beginning and end of the melt season.

From the statistical analysis of the large set of summer cyclones, we find that cyclone days in June and July have less local SIC loss and less region-wide SIA loss compared to days without cyclones. In contrast, cyclone days in August tend to be followed by slightly more SIC and SIA loss than non-cyclone days, although the relative enhancement of ice loss only persists for about one day. Although this analysis is limited to the Pacific sector of the Arctic, these results are broadly consistent with the localized effects

of cyclones on sea ice cover throughout the Arctic marginal ice zone (Finocchio et al. 2020, Schreiber and Serreze 2020).

The SW flux has the largest influence of all terms in the surface energy budget on both local and regional ice loss during the summer. In June and July, larger SW fluxes are associated with more short-term ice loss than smaller SW fluxes. Consequently, the largest reductions in local SIC in June tend to occur when skies are clear, indicating that the cloud radiative forcing is a key mechanism by which cyclones affect sea ice on short time scales early in the melt season. Calm and slightly cooler conditions in June also tend to be associated with larger local SIC loss, possibly because such conditions frequently accompany the clear skies that result in stronger SW fluxes at the surface. Later in the melt season, this correspondence between stronger SW fluxes and accelerated ice loss is only apparent for initially higher concentration sea ice. By August, cloudier conditions with more southerly surface winds are associated with slightly more local SIC loss. This suggests that the increased surface heat fluxes and enhanced sea ice advection in the warm sector of August cyclones enable these late-summer cyclones to be statistically associated with slightly enhanced local sea ice loss.

The advective tendency of SIC due to the 10-m wind is one of the most consistent predictors of both local and regional ice loss for the large sample of cyclones in the ECB region. We find the strongest relationship between advection and sea ice loss for low concentration sea ice in August. This supports previous studies arguing that the reduced mechanical strength of lower concentration sea ice makes it more susceptible to wind-induced drift and deformation (Hakkinen et al. 2008, Rampal et al. 2009, Spreen et al. 2011). The magnitude of the SIC advection within the ECB region is primarily controlled by the meridional component of the wind, with southerly winds favoring stronger negative sea ice advection. It is worth noting, however, that wind-driven advection only affects sea ice cover at a particular location, but does not affect the total sea ice volume across the Arctic.

We also analyzed two cyclones that reached a similar maximum intensity and affected a similar area in the East Siberian Sea, but at different times during the melt season. The early season cyclone that occurred in June 2012 primarily affected high concentration sea ice that extended from the central Arctic to the Siberian coast, while the late season cyclone that occurred in August 2016 affected an area of lower concentration sea ice located farther from the coast. In both cyclones, positive anomalies in the net surface energy flux are primarily a result of positive sensible heat flux anomalies, while negative anomalies in the net surface energy flux are due to negative SW flux anomalies. The magnitude and duration of positive anomalies in the net surface energy flux and surface wind speed are comparable between the two cyclones, yet the changes in the underlying sea ice distributions that followed the two cyclone cases are quite different. The August 2016 cyclone was followed by a reduction in sea ice area and a shift toward lower concentration sea ice (Fig. 26F). Meanwhile, there was only a temporary shift toward lower SIC and no change in sea ice area (Fig. 26E) following the June 2012 cyclone. This temporary shift to lower SIC may be an artifact of melt ponds being misclassified as open ocean (e.g. Meier et al. 2015), and this requires further investigation. Nonetheless, these apparent differences in the sea ice impacts of the two cyclones, despite their similar surface wind and energy forcing anomalies on sea ice, underscore the importance of the initial sea ice state and coupled upper ocean processes for determining the extent to which a cyclone impacts sea ice on short time scales (Lukovich et al. 2021).

A more detailed analysis of the surface energy balance and sea ice advection confirms many of the findings from the large set of cyclones regarding the evolving nature of how each forcing contributes to local sea ice changes during the melt season. For instance, SIC loss has a close relationship to the SW flux at each grid point during the June cyclone, and relatively weaker relationship to SW flux during the August cyclone. The sign of the relationship between SIC loss and the LW flux is the opposite of that between SIC loss and the SW flux, which is not surprising given the strong anti-correlation between SW and LW fluxes. The different magnitudes of each radiative flux in June versus August ultimately determine which flux is most influential in the net. In June, the SW fluxes are so strong and the modulation of SW fluxes by clouds so large that variability in the SW flux is the most influential term in the surface energy budget in terms of local sea ice loss. Specifically, the decreased SW flux at the surface due to clouds reflecting sunlight far outweighs the increased LW flux at the surface due to clouds in June,

such that clouds in June appear to behave more like an umbrella than a blanket (Perovich 2018). By August however, clouds behave like neither an umbrella nor a blanket in the net because the amount by which clouds decrease the SW flux at the surface, on average, is comparable to the amount by which they increase LW flux at the surface. As a result, the difference between clear and cloudy conditions is more likely to determine how much local SIC loss occurs during the June cyclone than during the August cyclone.

Although we generally find that larger surface heat fluxes are associated with enhanced ice loss in both the large sample of cyclones and for the two cyclone cases, this relationship is weaker than what was expected based on previous studies. Zhang et al. (2013) and Stern et al. (2020), for example, found that the anomalous sensible heat flux was the primary cause of the enhanced surface ice melt that occurred during the Great Arctic Cyclone of August 2012. The August 2016 cyclone analyzed in this study is comparable in many ways to the Great Arctic Cyclone, except that it was located further north and, perhaps as a result, was not associated with as much warm air advection from the Eurasian continent over sea ice. Based on our analysis in this study, the surface heat flux during the August 2016 cyclone was too small to explain the enhanced local SIC loss that followed. Instead, the SIC advection due to the surface wind forcing has a stronger statistical relationship to the local SIC loss following the August 2016 cyclone. This difference from the Great Arctic Cyclone in August 2012 highlights how the cyclone location relative to both land and the sea ice edge, and the underlying characteristics of the sea ice critically determine how a given cyclone will affect sea ice. Whether processes such as upper-ocean mixing, which was found to be a key cause of ice loss during the Great Arctic Cyclone, also contribute to the sea ice loss following the August 2016 cyclone requires further investigation. Fully coupled simulations of a diverse sample of Arctic summer cyclones spanning multiple decades would provide a richer understanding of the many ways in which cyclones affect sea ice, and could offer important clues as to how decadal variability and the rapidly evolving Arctic environment affect cyclone impacts on sea ice.

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