

UNIVERSITY OF OKLAHOMA

GRADUATE COLLEGE

TROPOPAUSE POLAR VORTEX LINKAGES TO ARCTIC AND MIDLATITUDE

PHENOMENA

A DISSERTATION

SUBMITTED TO THE GRADUATE FACULTY

in partial fulfillment of the requirements for the

Degree of

DOCTOR OF PHILOSOPHY

By

TOMER BURG
Norman, Oklahoma
2024

TROPOPAUSE POLAR VORTEX LINKAGES TO ARCTIC AND MIDLATITUDE
PHENOMENA

A DISSERTATION APPROVED FOR THE
SCHOOL OF METEOROLOGY

BY THE COMMITTEE CONSISTING OF

Dr. Steven Cavallo, Chair

Dr. David Parsons

Dr. Greg McFarquhar

Dr. Andrea Lang

Dr. Henry Neeman

Dedication

This work is dedicated to my grandparents. Without their love and support, none of this would have been possible.

Acknowledgments

The road to completing this PhD was not simple. I had known for many years that my passion was in meteorology, and that I wanted to learn more about what drives different weather phenomena and why they occur, but I did not know even a fraction of the possible career paths that can be pursued in this field. Even as an undergraduate student at the University at Albany, I often thought to myself that I was not capable of getting a PhD, especially coming from a family without a strong academic background. It was only after the passing of my grandpa who motivated me to pursue higher education, gaining substantial research experience at the Research Experience for Undergraduates at OU, and motivation from my family, Dr. Lance Bosart, and my undergraduate advisor Dr. Daniel Keyser, that I decided to give graduate school a try.

My graduate school experience has been a life changing one, to say the least. For many, their first day of graduate school is an exciting moment. My beloved grandma passed away merely days before starting graduate school. Processing the many life changes that had happened in a short time frame was a struggle, and I was fortunate to have the support of my family and closest friends. On the academic side, I learned invaluable knowledge from all the outstanding faculty, staff, and fellow students at the Department of Atmospheric and Environmental Sciences in the University at Albany, and especially from my MS advisor, Dr. Andrea Lang, and my co-advisors, Drs. Kristen Corbosiero and Ryan Torn.

After having spent 7 years at the same university, I reached out to Dr. Steven Cavallo with an interest to pursue Arctic meteorology at the School of Meteorology in OU. My time at OU was also not without challenges, between adjusting to living in a new part of the country far away from my family and most of my friends, and followed by a pandemic after which I relocated back home and worked remotely for the rest of my PhD. Even in spite of these challenges, I met amazing people in Norman who I am still fortunate to call my friends to this day, I gained new experiences and perspectives from completing my PhD studies at a different school, and am especially grateful to Steven and our AAARG research

group consisting of Matthew Bray, Joe Burzdak, Robin Clancy, Maddie Frank, Seraphine Hauser, Ipshita Majhi, Chen Nuo, and Chris Riedel, for their support along the way. I am also grateful to my PhD committee consisting of Steven and Drs. David Parsons, Greg McFarquhar, Andrea Lang, and Henry Neeman, for their contributions and feedback on this work.

I want to especially thank Drs. Bill Skamarock and Michael Duda at NCAR in Boulder, Colorado for hosting me as a summer visitor, and providing the computing resources on the Cheyenne and Derecho supercomputers to run the numerical simulations in this dissertation. Without their help and knowledge in teaching me more about the inner workings of numerical modeling, much of this work would not have been possible. Appreciation is also extended to former AAARG member Nick Szapiro for his help and support with adapting his TPV modification methodology for this work. I also want to acknowledge Ariel Cohen for his encouragement and support along the way, and my colleagues and friends, including Massey Bartolini, Noah Brauer, Hristo Chipilski, Simon Lee, Kevin Lupo, and Brendan Wallace, among many others.

Table of Contents

Dedication	iv
Acknowledgments	v
List of Tables	x
List of Figures	xi
Abstract	xxiv
1 Introduction and Background	1
1.1 Introduction to TPVs	3
1.2 TPV history in literature	6
1.2.1 Arctic tropopause folds	6
1.2.2 Early identification of tropopause-based vortices	7
1.2.3 Modern analyses of TPVs	9
1.3 TPV characteristics and dynamics	10
1.3.1 TPV characteristics	11
1.3.2 TPV dynamics	13
1.3.3 TPV preference in high latitudes	18
1.4 Tropopause definitions	19
1.5 TPV remote dynamics	20
1.5.1 TPV-Arctic cyclone linkages	20
1.5.2 TPV-polar low linkages	24
1.5.3 TPV-cold air outbreak linkages	27
1.6 Outline and Objectives	29
1.6.1 Systematic examination of rapidly deepening ACs	30
1.6.2 Quantifying TPV intensity linkage to cyclogenesis	32
1.6.3 February 2021 CAO linkage to TPVs	33
1.6.4 Dissertation outline	34
2 Data and methods	43
2.1 Data sources	43
2.1.1 Calculations	44
2.1.2 Ensemble sensitivity analysis	45
2.1.3 Cyclone tracking and identification overview	46
2.1.4 TPV identification and tracking	47
2.2 Rapidly deepening ACs	48
2.2.1 Review of rapid deepening criteria	48

2.2.2	Criteria for rapidly deepening ACs	49
2.2.3	Storm-centered composites	51
2.2.4	TPV matching	52
2.3	Model experiment setup	53
2.3.1	TPV intensity modification algorithm	54
2.3.2	August 2012 AC experiment setup	56
2.3.3	February 1987 PL experiment setup	57
2.3.4	February 2021 CAO experiment setup	58
3	Characteristics of rapidly deepening Arctic cyclones	71
3.1	Rapidly deepening AC climatology	72
3.2	Summer and winter composites	74
3.2.1	TPV proximity comparison	75
3.2.2	EGR comparison	76
3.3	Regional composites	77
3.4	Onset of rapid deepening	79
3.5	Rapidly deepening vs. ordinary ACs	80
3.6	Discussion	81
4	TPV linkages to polar low and Arctic cyclone cases	98
4.1	August 2012 great Arctic cyclone	98
4.1.1	ERA5 and control simulation overview	99
4.1.2	TPV modification simulations	100
4.2	February 1987 polar low	103
4.2.1	ERA5 and control simulations overview	103
4.2.2	IC uncertainty ensemble	105
4.3	Discussion	107
5	TPV linkages to the February 2021 cold air outbreak	123
5.1	Case overview	124
5.1.1	Large-scale precursors	124
5.1.2	TPV and CAO evolution	125
5.1.3	Air parcel trajectory analysis	127
5.2	MPAS simulation experiments	129
5.2.1	Synoptic overview of simulations	129
5.2.2	southern Great Plains temperature analysis	133
5.3	Discussion	134
5.3.1	Overview of additional simulations	137
6	Conclusions	162
6.1	Recommendations for future work	170

List of Tables

1.1	Summary of experiments from Cavallo and Hakim (2013), denoting inclusion of longwave radiation (LW), water vapor (WV), shortwave radiation (SW), and clouds.	17
2.1	MPAS v8.0.1 model configuration used for all simulations.	54

List of Figures

- 1.1 Northwest–southeast oriented cross sections through an Icelandic low on 12 April 1983 using two datasets for approximately the same analysis time. (a) ECMWF analysis, with isentropes in thin black contours every 10 K and PV in thick black contours. The thin lines indicate isentropes at 10 K intervals and the thick lines PV at 1 unit intervals. The 0.75 PVU unit contour ($1 \text{ PVU} = 1 \times 10^{-6} \text{ K kg}^{-1} \text{ s}^{-1} \text{ m}^2$) is shown by a heavy dashed line. (b) In-situ observations from aircraft and dropwindsondes, with isentropes in thin black contours every 2 K, isotachs in dashed lines every 10 m s^{-1} , and estimated tropopause in solid black line. Figure reproduced from Hoskins et al. (1985), their Fig. 9. 36
- 1.2 Cross section through an isolated upper-tropospheric (a) cyclonic and (b) anticyclonic PV anomaly. The location of the anomaly is denoted by stippled regions. The thick black contour represents the tropopause, and the two sets of thin contours denote potential temperature every 5 K and transverse velocity every 3 m s^{-1} . Figure reproduced from Hoskins et al. (1985), their Fig. 15. 37
- 1.3 Composite TPV west–east cross-vortex sections of the anomalous (a) temperature (K), (b) v-wind component (m s^{-1}), (c) Ertel potential vorticity (PVU), and (d) relative humidity (%). The thick, solid (dashed) black contour is the composite (background) tropopause. The 0 contour is denoted by the thin, solid contour. Figure and caption reproduced from Cavallo and Hakim (2010), their Fig. 9. 38

1.4	Composite TPV west–east cross-vortex sections of the anomalous EPV tendency due to (a) all diabatic effects, (b) radiation, (c) latent heating, and (d) all diabatic effects except for radiation and latent heating. Units are PVU day ^{−1} and the contour interval is 0.04 PVU day ^{−1} . The thick, solid (dashed) black contour is the composite (background) tropopause. The 0 contour is denoted by the thin, solid contour. Figure and caption reproduced from Cavallo and Hakim (2010), their Fig. 9.	39
1.5	Time series plots of vortex amplitude for experiments E1–E5 as labeled in the right margin. Amplitudes are computed on the 2-PVU surface. Figure reproduced from Cavallo and Hakim (2013), their Fig. 3.	40
1.6	Occurrence frequency (%) of (a) all TPVs, (b) all cold pools, (c) TPVs transported to middle latitudes (equatorward of 60°N), and (d) cold pools transported to middle latitudes (equatorward of 60°N). Figure reproduced from Biernat et al. (2021), their Fig. 7.	41
1.7	Backward trajectories initiated at 1200 UTC 30 Jan 2019, with dots plotted at 1-h intervals and colors representative of the parcel potential temperature. Figure reproduced from Lillo et al. (2021), their Fig. 5.	42
2.1	Number of TPV occurrences within 555 km of a point using (a, b) criteria from Cavallo and Hakim (2009), and (c, d) criteria from Szapiro and Cavallo (2018). Plotted are the (a, c) Northern Hemisphere, and (b, d) Southern Hemisphere.	61
2.2	Bergeron criteria as a function of latitude in the North Hemisphere. Blue line denotes a value of 1.0 Bergeron. Red lines denote the default 60°N reference latitude and associated 24 hPa per 24 h threshold.	62
2.3	500-km track density of the portion of cyclone tracks where the 24 h deepening rate exceeds 1 Bergeron, using 1984–2016 ERAI cyclone tracks. . . .	63

2.4	North Hemisphere comparisons of zonally averaged (a) Bergeron criteria (dashed line) and 95th percentile of observed cyclone deepening rates per 2.5° latitude bin (solid line), (b) annual mean meridional gradient of zonal mean 1000–500 hPa thickness (dam deg^{-1}), and (c) annual mean 850-hPa EGR (day^{-1}).	64
2.5	Same as Fig. 2.3, but for the 95th percentile threshold of observed cyclone deepening rates.	65
2.6	MPAS regional domain used for the 2012 “Great Arctic Cyclone” simulations. Within the domain, ERA5 ICs are shown for 2 PVU potential temperature (K, fill) and MSLP (every 4 hPa, brown contours). Mesh spacing within this domain is approximately 15 km.	66
2.7	MPAS regional domain used for the February 1987 PL simulations. Within the domain, ERA5 ICs are shown for 2 PVU potential temperature (K, fill). The solid white line denotes the ERA5 PL track based on hourly MSLP minima, and the thin yellow line denotes the ERA5 TPV track based on hourly 2 PVU potential temperature minima. Mesh spacing within this domain is approximately 15 km.	67
2.8	MPAS regional domain used for the February 2021 CAO simulations. Mesh spacing within this domain is approximately 15 km.	68
2.9	Domain within which the TPV intensity modification algorithm is applied to as a function of time. The temporal adjustments in this domain encompass both TPVs of interest, while avoiding a separate PV lobe to the southwest and attempting to separate one of the TPVs of interest from a fragment that splits into a separate TPV.	69
2.10	Southern Great Plains domain used for several analyses in Chapter 5.	70

3.1	Occurrence frequency (%) of (a) all 6 h cyclone positions where RD criteria is met, and (b) cyclogenesis locations of all cyclones that meet RD criteria. Occurrence frequency is calculated as the percent of 6-h time steps between 1979–2022 where a grid point is located within 500 km of the portion of the track of an AC where the criteria specified above is met. The Arctic Circle is denoted by the dotted blue line.	83
3.2	Top row shows annual Arctic-wide composite mean of 2 PVU potential temperature (K, green contours), normalized anomaly (sigma, fill), and MSLP (hPa, black contour) at the (a) start, (b) peak, and (c) end of the rapid deepening episode. Bottom row shows the percent of TPV basins in a gridpoint out of all basins for the closest TPV to the AC, at the (d) start, (e) peak, and (f) end of the rapid deepening episode.	84
3.3	Difference in composite mean (a) 500-hPa geopotential height and (b) 2 PVU potential temperature normalized anomalies at the peak minus onset of the rapid deepening episode. Contours show the average of the start and peak composites for (a) 500-hPa geopotential height (dam), and (b) 2 PVU potential temperature (K). Dots denote differences statistically significant at the 99% level.	85
3.4	Same as Fig. 3.1, but for locations of rapidly deepening ACs for (a) annual cases, (b) summer cases, and (c) winter cases. Bottom row shows cyclogenesis location density for (d) annual cases, (e) summer cases, and (f) winter cases.	86
3.5	Same as top row of Fig. 3.2, for annual cases, showing the (a) start, (b) peak, and (c) end of the RD episode. Second row (d-f) shows the same for summer cases, and the third row (g-i) shows the same for winter cases. . . .	87

3.6	Box and whisker plot for the distribution of AC distance from the nearest TPV mass centroid for the start, peak, and end of the RD episode between annual, summer, and winter cases. Blue stars denote differences between the summer and winter distributions are statistically significant at the 99% level, and white stars show statistical significance at the 95% level.	88
3.7	Difference in occurrence frequency for cyclone location density at the peak of the RD episode, for cases with an AC-TPV distance below the lower quartile minus cases with an AC-TPV distance above the upper quartile. (a) shows this difference for summer cases, and (b) shows this difference for winter cases.	89
3.8	Difference in occurrence frequency for cyclone location density at the peak of the RD episode, for (top row) cases with EGR above the upper quartile minus cases with EGR below the lower quartile, for (a) annual cases, (b) summer cases, and (c) winter cases. Bottom row shows the same as the top row, but for EGR normalized anomaly.	90
3.9	Box and whisker plot for the distribution of EGR normalized anomaly for the start, peak, and end of the RD episode between annual, summer, and winter cases. Blue stars denote differences between the start and peak, or peak and end, distributions are statistically significant at the 99% level, and white stars show statistical significance at the 95% level. No stars indicate no significant differences. The black arrows highlight the differing trend in composite EGR normalized anomaly, most prominent in the upper quartiles, throughout the RD episode between the summer and winter cases.	91
3.10	Correlation between EGR normalized anomaly and distance from the nearest TPV for (a) summer cases, and (b) winter cases. The Pearson r^2 value and p-value are displayed in the upper left corner, with blue denoting the correlation is statistically significant.	92

3.11	Same as Fig. 3.4, but for top row showing RD density for (a) Pacific cases and (b) Atlantic cases, and bottom row showing cyclogenesis location density for (c) Atlantic cases, and (d) Pacific cases.	93
3.12	Box and whisker plot for the distribution in EGR normalized anomaly for the start, peak, and end of the RD episode between Atlantic and Pacific cases. Blue stars denote differences between Atlantic and Pacific cases are statistically significant at the 99% level, and white stars show statistical significance at the 95% level. No stars indicate no significant differences.	94
3.13	Difference in composite mean (a) MSLP and (b) 2 PVU potential temperature normalized anomalies for cases whose RD coincides with cyclogenesis minus cases whose RD onset occurs at least 24 h or longer after cyclogenesis. Contours show the average of the start and peak composites for (a) MSLP (dam), and (b) 2 PVU potential temperature (K). Dots denote differences statistically significant at the 99% level.	95
3.14	Occurrence frequency for cyclone location density at the onset of the RD episode for cases whose RD onset coincides with cyclogenesis minus cases whose RD onset occurs at least 24 h or longer after cyclogenesis, for (a) annual cases, and (b) summer cases. Red shading denotes locations where cyclones in the onset category are more frequent, and blue shading denotes locations where cyclones in the 24 hour category are more frequent.	96
3.15	Box and whisker plot comparing distributions between rapidly deepening and ordinary ACs for (a) distance from the nearest TPV, (b) EGR, and (c) EGR normalized anomaly. Blue stars denote differences between Atlantic and Pacific cases are statistically significant at the 99% level, and white stars show statistical significance at the 95% level. No stars indicate no significant differences.	96

3.16	Composite mean of QG omega term A for (a) annual, (b) summer, and (c) winter cases, scaled by $10^{12} \text{ kg m}^{-3} \text{ s}^{-3}$	97
4.1	ERA5 2 PVU potential temperature (K) in fill and MSLP below 1008 hPa in contours. The locations of TPV 1 and TPV 2 are highlighted where applicable. Frames are valid at (a) 0000 UTC 3 August 2012, (b) 0000 UTC 4 August 2012, (c) 0000 UTC 5 August 2012, (d) 1200 UTC 5 August 2012, (e) 0000 UTC 6 August 2012, and (f) 1200 UTC 6 August 2012. . . .	110
4.2	MPAS control simulation 2 PVU potential temperature (K) in fill and MSLP below 1008 hPa in contours. The locations of TPV 1 and TPV 2 are highlighted where applicable. Frames are valid at (a) 0000 UTC 4 August 2012, (b) 0000 UTC 5 August 2012, (c) 1200 UTC 5 August 2012, (d) 0000 UTC 6 August 2012, (e) 1200 UTC 6 August 2012, and (f) 0000 UTC 7 August 2012.	111
4.3	Comparison between ERA5 (black) and the control simulation (red) for AC12 cyclone track (left) and minimum MSLP (hPa, right).	111
4.4	MPAS azimuthally averaged cross section centered at the core of TPV 1 at the end of the 24 h TPV modification window, valid 0000 UTC 5 August 2012, for (a) the control simulation, and (b) the TPV1++ simulation. EPV (PVU) is shown in fill, azimuthally averaged wind speed (knots) are in blue contours, and the solid black line denotes the azimuthally averaged tropopause level from the MPAS output.	112

4.5	(Left) Comparison of AC12 cyclone track between the control simulation (black) and TPV1++ simulation (light blue), with dots every 24 h beginning 0000 UTC 4 August 2012. (Right) Control simulation 3-h MSLP change (hPa, dotted line), difference in the TPV1++ simulation minus the control simulation for term A of the QG omega equation between 500–700 hPa averaged within 4 hPa of the minimum MSLP of AC12 ($10^{13} \text{ kg m}^{-3} \text{ s}^{-3}$, blue line), 700-hPa geostrophic temperature advection averaged within 4 hPa of the minimum MSLP of AC12 (10^5 K s^{-1} , red line), and 3-h MSLP change (hPa, black).	113
4.6	Same as Fig. 4.5, but for the TPV2++ simulation.	113
4.7	500-hPa geostrophic absolute vorticity advection by the geostrophic wind ($\times 10^{10} \text{ s}^{-2}$, fill), smoothed 500-hPa geopotential height (dam, black contours), and MSLP below 996 hPa (hPa, purple contours) for (a) control simulation, and (b) Trough- simulation, at forecast hour 51 valid 0300 UTC 6 August 2012.	114
4.8	Difference in the Trough- simulation minus the control simulation for term A of the QG omega equation between 500–700 hPa averaged within 4 hPa of the minimum MSLP of AC12 ($10^{13} \text{ kg m}^{-3} \text{ s}^{-3}$, blue line), 700-hPa geostrophic temperature advection averaged within 4 hPa of the minimum MSLP of AC12 (10^5 K s^{-1} , red line), and 3-h MSLP change (hPa, black).	115
4.9	(a) AC12 minimum MSLP (hPa), and (b) 12-h change in AC12 minimum MSLP (hPa), for the control simulation (black), TPV1++ simulation (light blue), TPV2++ simulation (dark blue), Both++ simulation (purple), and Trough- simulation (red).	116
4.10	AC12 simulated track based on minimum MSLP location for the control simulation (black), TPV1++ simulation (light blue), TPV2++ simulation (dark blue), Both++ simulation (purple), and Trough- simulation (red).	117

4.11	ERA5 2 PVU potential temperature (K) in fill and MSLP (hPa) in contours. Frames are valid at (a) 0000 UTC 25 February 1987, (b) 1200 UTC 25 February 1987, (c) 0000 UTC 26 February 1987, (d) 1200 UTC 26 February 1987, (e) 0000 UTC 27 February 1987, and (f) 1200 UTC 27 February 1987.	118
4.12	Control simulation cyclone tracks and minimum MSLP (hPa) from MPAS control simulations initialized on 0600 UTC (light blue) and 1200 UTC (dark blue) 25 February 1987. Thicker lines denote lower MSLP, and thinner lines denote higher MSLP.	119
4.13	Ensemble 75 km track density (%) for the (a) 0600 UTC 25 February initialization, and (b) 1200 UTC 25 February initialization.	120
4.14	ESA analysis for the ensemble initialized 0600 UTC 25 February 1987. (a) Ensemble mean MSLP (hPa) valid 1200 UTC 26 February 1987, with the leading EOF shown in contours. (b) Sensitivity (fill) of the leading EOF in the domain shown in (a), valid at 1200 UTC 26 February 1987, to 2 PVU potential temperature (ensemble mean in contour) valid 0900 UTC 25 February 1987.	120
4.15	(a) 75 km track density difference (%) between the stronger TPV ensemble minus the control ensemble initialized 0600 UTC 25 February 1987. (b) PL87 minimum MSLP difference (hPa) between the stronger TPV ensemble minus the control ensemble.	121
4.16	(a) 75 km track density difference (%) between the weaker TPV ensemble minus the control ensemble initialized 1200 UTC 25 February 1987. (b) PL87 minimum MSLP difference (hPa) between the weaker TPV ensemble minus the control ensemble.	121

4.17	Idealized schematic representing the impact of intensifying a TPV embedded within a synoptic-scale trough, versus weakening the amplitude of the synoptic-scale trough, on the track of a developing downstream surface cyclone.	122
5.1	Difference in TPV probability per day, defined as the average probability of at least one TPV being located within 555 km of the corresponding point per day over the respective period (i.e., DJF or JJA), for winter minus summer months in the (a) Northern Hemisphere, where DJF denotes winter and JJA denotes summer, and (b) Southern Hemisphere, where JJA denotes winter and DJF denotes summer.	139
5.2	Minimum 2-meter temperatures (°F) between 0000 UTC 17 February 2021 and 2300 UTC 17 February 2021, as derived from the hourly 2.5 km horizontal resolution Real Time Mesoscale Analysis (RTMA; De Pondeca et al. 2011).	140
5.3	ERA5 2 PVU potential temperature (K) in fill and mean sea level pressure (hPa) in contours. Top row shows the evolution of the Arctic anticyclone, valid at (a) 0000 UTC 1 February 2021, (b) 0000 UTC 3 February 2021, and (c) 0000 UTC 5 February 2021. Bottom row shows the evolution of the Greenland anticyclone, valid at (d) 0000 UTC 5 February 2021, (e) 0000 UTC 6 February 2021, and (f) 0000 UTC 7 February 2021.	141
5.4	ERA5 2 PVU potential temperature (K) in fill and mean sea level pressure (hPa) in contours. Frames are valid at (a) 0000 UTC 8 February 2021, (b) 0000 UTC 10 February 2021, (c) 0000 UTC 12 February 2021, (d) 0000 UTC 13 February 2021, (e) 0000 UTC 14 February 2021, and (f) 0000 UTC 15 February 2021.	142

5.5	Time series of ERA5 2-meter temperatures averaged over the denoted southern Great Plains domain from February 2021 (black line), compared against ERA5 1950–2022 average (blue line), interquartile range (blue shading), and minimum record excluding 2021 (dashed blue line).	143
5.6	ERA5 Cross section of EPV (PVU; fill), potential temperature (K; black contours) and wind speed isotachs (knots; blue contours), valid at 1200 UTC 12 February 2021. Upper right inset shows 2 PVU potential temperature valid at the same time.	144
5.7	Backward air parcel trajectories ending in Oklahoma City at 500 m altitude on (a) 1200 UTC 15 February 2021, and (b) 1200 UTC 16 February 2021. Images generated from the HYSPLIT program (Stein et al. 2015).	145
5.8	Backward parcel trajectories ending in Oklahoma City on (a, b) 1200 UTC 15 February 2021 in the top row, and (c, d) 1200 UTC 16 February 2021 in the bottom row, all ending at 500 m altitude. Parcel coordinates relative to TPV B1 are shown on the left (a, c), and TPV A on the right (b, d). Colors denote time relative to the end of the parcel trajectory, and line widths denote the fraction of the pressure of the parcel between the surface pressure and 2 PVU pressure at the center of the TPV, such that thicker lines denote a parcel altitude near the tropopause level at the vortex core, and thinner lines denote a parcel altitude near the surface.	146
5.9	Backward air parcel trajectory ensembles ending in Oklahoma City on 1200 UTC 15 February 2021, at 250 m altitude on the left (a, c) and 500 m altitude on the right (b, d). Parcel coordinates relative to TPV B1 are shown on the top (a, b), and TPV A on the bottom (c, d). Colors and line thickness follow the same convention as Fig. 5.8.	147
5.10	Same as Fig. 5.9, but ending at 1200 UTC 16 February 2021.	148

5.11	Percent of HYSPLIT ensemble members ending at 1200 UTC 15 February 2021 at 250 m altitude above ground level (cyan), 500 m altitude (light blue), and 750 m altitude (dark blue), whose parcel trajectories are closest to TPV B1 (solid lines) or TPV A (dashed lines) and within 1000 km of that TPV.	149
5.12	Same as Fig. 5.11, but ending at 1200 UTC 16 February 2021.	150
5.13	Control simulation 2 PVU potential temperature (K), valid at (a) 0000 UTC 8 February 2021, (b) 0000 UTC 10 February 2021, (c) 0000 UTC 12 February 2021, (d) 0000 UTC 13 February 2021, (e) 0000 UTC 14 February 2021, and (f) 0000 UTC 15 February 2021.	151
5.14	2 PVU potential temperature (K) valid at 1200 UTC 9 February 2021, at the end of the TPV modification period compared across all simulations. The control simulation is shown in the center.	152
5.15	Same as Fig. 5.14, but valid at 1200 UTC 11 February 2021. TPV B1 is located to the south, and TPV A to the north, in all panels.	153
5.16	Same as Fig. 5.14, but valid at 0000 UTC 14 February 2021.	154
5.17	MPAS ensemble standard deviation in 500-hPa geopotential height (dam) in fill, and control member 500-hPa geopotential height (dam) in contours, valid (a) 1200 UTC 9 February 2021, (b) 0000 UTC 11 February 2021, (c) 0000 UTC 12 February 2021, (d) 0000 UTC 13 February 2021, (e) 0000 UTC 14 February 2021, and (f) 0000 UTC 15 February 2021.	155
5.18	MPAS ensemble standard deviation in 2 PVU potential temperature (K) in fill, and control member 2 PVU potential temperature (K) contoured every 5 K up to 290 K, valid (a) 1200 UTC 9 February 2021, (b) 0000 UTC 11 February 2021, (c) 0000 UTC 12 February 2021, (d) 0000 UTC 13 February 2021, (e) 0000 UTC 14 February 2021, and (f) 0000 UTC 15 February 2021.	156

5.19	2-meter temperature ($^{\circ}\text{C}$) averaged over the southern Great Plains domain for ERA5 (dotted black line) and the various MPAS simulations performed for this experiment.	157
5.20	Vertical temperature (red) and dewpoint (blue) profiles for (a) HRRR and (b) MPAS control simulation in solid lines, with radiosonde observations in transparent colors. Both simulations were initialized at 0000 UTC 8 February 2021, valid at 1200 UTC 9 February 2021.	157
5.21	Vertical (red) and dewpoint (blue) profiles for the MPAS control simulation in solid lines, with radiosonde observations in transparent color, initialized at 0000 UTC 8 February 2021 and valid at 1200 UTC 11 February 2021. . .	158
5.22	2-meter temperature bias ($^{\circ}\text{C}$) from the MPAS control simulation relative to RTMA at forecast hour 66, valid at 1800 UTC 10 February 2021. . . .	159
5.23	Sensitivity of 2-meter temperatures averaged over the southern Great Plains at 0000 UTC 15 February 2021 to 500-hPa geopotential height valid 0600 UTC 11 February 2021.	159
5.24	Same as Fig. 5.23, but showing sensitivity to 2 PVU potential temperature.	160
5.25	Schematic showing the general relationship between TPV intensity and cold pool intensity (blue line) contrasting with a different evolution of the PV lobe over western Canada in simulations with a more intense TPV than the control simulation leading to a poleward displacement of the core of the cold pool (red line). The end result is the control simulation producing the coldest temperatures in the southern Great Plains (black line).	160
5.26	Same as Fig. 5.19, but for the MPAS experiments that did not incorporate a moving domain for the TPV modification region that filters out TPV B2 and a PV lobe over southwest Canada.	161

Abstract

Tropopause polar vortices (TPVs) are coherent, closed tropopause-based vortices that spend at least 60% of their lifetime poleward of 65° latitude. TPVs are identified by a local minima in potential temperature and height, and a local maxima in potential vorticity on the dynamic tropopause. TPVs are most common in the Arctic, where they are often associated with the intensification of Arctic cyclones (ACs), but on occasion exit the Arctic into the midlatitudes where they are often associated with cold air outbreaks (CAOs).

This dissertation investigates TPV linkages to ACs, CAOs, and polar lows (PLs), focusing on systematic TPV-AC linkages and case studies of TPV intensity linkages to an AC case, a PL case, and a major CAO. Rapidly deepening ACs are commonly associated with an upstream TPV which becomes vertically aligned with the AC by the end of the rapid deepening episode. Summer cases are associated with closer proximity to the closest TPV and less lower-tropospheric baroclinic instability than winter cases, with summer cases often over the central Arctic Ocean and winter cases often in the North Atlantic into the Barents and Kara seas. Additionally, ACs whose rapid deepening episode coincides with cyclogenesis are less likely to be over the central Arctic Ocean and are farther from the closest TPV relative to ACs that develop at least 24 h prior to the onset of rapid deepening.

Numerical simulations designed to modify the intensity of two TPVs associated with the August 2012 “Great Arctic Cyclone” show that stronger TPVs are associated with a faster peak deepening rate and earlier peak intensity of the cyclone, but with the closer TPV to the AC exhibiting a greater impact on the intensification rate of the AC, and the farther TPV exhibiting a greater impact on the track of the AC. The increased intensification rate is primarily associated with stronger differential cyclonic vorticity advection downstream of the TPVs. Weakening the broader upper-tropospheric trough within which the closer TPV is embedded in results in a substantially weaker and more progressive AC. Applying the same TPV modification methodology to an intense PL case only results in

very minimal impact on the intensity of the PL, and a 10-member initial condition uncertainty ensemble shows the track and intensity of the PL are more sensitive to the amplitude of a ridge upstream of the TPV than the intensity of the TPV.

Finally, an investigation of the role of two merging TPVs in a historic CAO in the southern Great Plains in February 2021 shows both TPVs had a direct role in the evolution of the CAO, but that backward air parcel trajectories from the southern Great Plains were mostly associated with one TPV instead being distributed equally between both TPVs. The control simulation produces the coldest temperatures at the peak of the CAO in the southern Great Plains, as intensifying both TPVs results in an earlier merger and stronger cold pool but a slower and farther north resultant TPV, while weakening both TPVs results in a weaker cold pool. This result differs from past studies of TPV-CAO linkages where a single, fast-moving TPV exhibited a more direct linkage between TPV intensity and CAO magnitude.

Chapter 1

Introduction and Background

A typical first impression of the Arctic might be of a frigid, desolate, inhospitable space far from civilization. However, it is anything but desolate from a scientific and meteorological perspective. The Arctic is a fundamental component of global meteorology and climate, and is associated with a vibrant range of meteorological phenomena. From a socio-economic perspective, the Arctic is home to approximately 7 million people, a figure which is projected to increase to over 10 million by 2080 (Heleniak 2020). Some of the Arctic population resides in coastal settlements, which are especially vulnerable to erosion and flooding due to sea level rise (e.g., Nielsen et al. 2022), and an abundance of oil and natural gas reserves in the Arctic make it of economic importance (e.g., Petrick et al. 2017).

Solar radiation plays an important role in Arctic meteorology. The Arctic is commonly defined as the region poleward of the Arctic Circle, which is located at approximately 66° latitude. Poleward of this latitude, and especially closer to the pole, there is little to no solar heating during the winter as the sun does not rise, while during the summer, the sun does not set. The resulting radiative imbalance during the winter leads to the development of a large hemispheric-scale vortex in the stratosphere, often centered near the pole, known as the stratospheric polar vortex (SPV), which forms in the fall and weakens in the spring (Waugh et al. 2017). On occasion, tropospheric planetary waves propagating upward into the stratosphere can disrupt and weaken the SPV, resulting in what is known as a sudden stratospheric warming (SSW) event, which can have substantial impacts on midlatitude weather pattern in subsequent weeks (Butler et al. 2017). It is in part for this reason that the Arctic stratosphere is recognized as a contributor to subseasonal-to-seasonal (S2S) range predictability (e.g., Merryfield et al. 2020).

The Arctic is also recognized as a region of relatively poor forecast skill. Global numerical weather prediction (NWP) skill scores are routinely lower in the Arctic than in lower latitudes (e.g., Riedel 2020; Capute and Torn 2021). Owing to the sparse population and largely inhospitable climate, in-situ data observations are sparse as well, which increases the importance of observations that do exist; in one case study, simply denying three well-situated radiosondes substantially degraded forecast skill of an Arctic cyclone (AC) and associated tropopause-based vortex (Johnson and Wang 2021). Arctic forecast skill has important implications on midlatitude predictability; removing Arctic observations significantly degrades synoptic forecast skill in the midlatitudes (Sato et al. 2018; Day et al. 2019), while additional Arctic observations can improve forecast skill of lower latitude phenomena such as tropical cyclones (Sato et al. 2018).

This dissertation focuses predominantly on tropopause polar vortices (TPVs) (e.g., Cavallo and Hakim 2009, 2010, 2012, 2013), which are another phenomenon common in the Arctic and which can have substantial impacts on various phenomena within and outside of the Arctic. The overarching goal of this dissertation is to review our historical and current understanding of TPVs, including their characteristics and dynamics, and quantify their linkages to various phenomena such as ACs, cold air outbreaks (CAOs), and Polar lows (PLs). Specifically, their systematic linkages to rapidly intensifying ACs will be evaluated, with several case studies to investigate their role in the development of an AC, polar low (PL), and CAO. The remainder of Chapter 1 discusses a literature review of TPVs, their characteristics and dynamics, linkages to various Arctic and midlatitude phenomena, and concludes with the objectives of this dissertation and underlying hypotheses.

1.1 Introduction to TPVs

TPVs¹ are coherent vortices in the tropopause layer, identified by a local minimum in potential temperature on the dynamic tropopause (Hakim and Canavan 2005; Cavallo and Hakim 2009). TPVs are generally subsynoptic scale features with radii most often ranging from 100 to 1000 km and lifetimes ranging from several days to several months (Hakim 2000; Hakim and Canavan 2005).

The term “TPV” was motivated by the desire to study the upper-level precursor disturbances that induce surface cyclones and produce organized vertical motion in extratropical regions. Hakim (2000) showed that these upper-level disturbances were, in fact, coherent vortices due to the nonlinear dynamics that better characterize higher latitudes. Subsequently, these vortices were referred to as coherent tropopause disturbances (CTDs) (Pyle et al. 2004). However, given that CTDs often originate from higher latitudes in isolation from the jet stream, Hakim and Canavan (2005) imposed further restrictions, requiring that these features have lifetimes of at least two days and that at least 60% of their lifetime is spent poleward of 65° latitude. These restrictions were imposed to study the vortices in isolation from the jet stream before additional complicating dynamics play a role in their subsequent evolution. Cavallo and Hakim (2009) adopted these criteria to define TPVs. TPVs are distinguishable from the similarly named tropospheric polar vortex and from the stratospheric polar vortex as TPVs are of a much smaller spatial scale, often occur on shorter temporal scales, and can exist within the stratospheric polar vortex and tropospheric polar vortex (Vaugh et al. 2017). Furthermore, their governing nonlinear dynamics distinguish them from quasi-linear Rossby waves (RWs) (Hakim 2000).

Studies of TPVs in isolation from the jet stream yielded significant progress in the understanding of their maintenance and intensification. Cavallo and Hakim (2009) used a potential vorticity (PV) tendency diagnostic approach to first quantify the role that cloud-top

¹TPVs are commonly used to refer to cyclonic vortices, and references to TPVs in this dissertation will follow this convention unless otherwise noted.

radiative cooling plays in strengthening a long-lived TPV. Soon thereafter, using a sample of 568 TPVs in northern Canada derived from a long-term numerical model simulation, Cavallo and Hakim (2010) found that diabatic heating due to longwave (LW) radiative cooling is the primary process responsible for the strengthening of TPVs. Furthermore, they found that latent heating plays a destructive role in TPV intensification. While latent heating has a significant impact on TPV intensity, latent heating rates in high latitudes are much smaller than the LW radiative cooling rates that strengthen TPVs, resulting in the net dominance of radiation over latent heating in high latitudes. This dominance may be used to explain the observed longevity of TPVs.

To more clearly isolate the processes responsible for TPV intensification, Cavallo and Hakim (2013) performed an idealized modeling study based on the observed structure of TPVs. As stated by Cavallo and Hakim (2013), LW radiative cooling is stronger where water vapor decreases with height more rapidly than in surroundings. In clear-sky conditions, LW radiative cooling is strongest in the vortex core, just below the tropopause, where the lower tropopause height relative to the surrounding environment is associated with a downward intrusion of dry stratospheric air. This downward intrusion establishes a narrow vertical transition zone between the dry stratospheric air and moist tropospheric air, and accordingly a more rapid WV decrease with height relative to the surrounding environment. Subsequently, EPV creation is maximized below the vortex core near the tropopause, where the vertical gradient of LW radiative cooling is largest. In cloudy conditions, cloud-top LW radiative cooling further enhances EPV creation and TPV intensification.

Given the linkages between TPVs and numerous Arctic and midlatitude weather phenomena and processes, an improved understanding of TPVs is important from a predictability standpoint. TPVs are often precursors to midlatitude cyclones (e.g., Uccellini et al. 1985; Sanders 1986; Bosart et al. 1996). Therefore, understanding the processes that control TPV evolution can result in improved predictions of midlatitude cyclones and their

associated impacts. The formation and intensification of midlatitude cyclones often involve interactions between TPVs and the midlatitude jet stream. These interactions render the original latitudinal criteria for TPVs proposed by Hakim and Canavan (2005), which were selected to study TPVs in isolation from the midlatitude jet stream, overly restrictive when studying portions of TPV life cycles associated with midlatitude cyclogenesis. The reason the original latitudinal criteria are overly restrictive is that TPVs associated with midlatitude cyclogenesis may originate in the Arctic yet spend a large portion of their life cycles outside of the Arctic. Szapiro and Cavallo (2018) relaxed the requirement that TPVs spend 60% of their lifetimes poleward of 65° latitude to instead require they form poleward of 60° latitude. This refinement captures both the TPV presence in the Arctic as well as their transportation into the midlatitudes. This relaxation of the definition provided an opportunity to revisit the impacts a TPV may have in lower latitudes. For example, recent TPV studies focused on linkages with extreme CAOs (Papritz et al. 2019; Biernat et al. 2021; Lillo et al. 2021) and tornado outbreaks in the continental United States (Bray et al. 2021).

Meanwhile, the impacts of TPVs before they exit the Arctic continue to be a focus of active research. In particular, some of the rapid reductions in Arctic sea ice have been noted to occur in association with ACs (e.g., Simmonds and Rudeva 2012; Zhang et al. 2013), and TPVs have been noted to play important roles in AC development and evolution (e.g., Yamagami et al. 2017; Gray et al. 2021). The reductions in Arctic sea ice enable shipping in the Arctic due to increased accessibility and shorter shipping routes (e.g., Mudryk et al. 2021; Li and Lynch 2023), and has also led to severe coastal erosion issues that pose the necessity of relocating coastal communities (e.g., Barnhart et al. 2014; Fang et al. 2018).

As documentation of TPVs in the literature is relatively recent, the foregoing discussion motivates a review of historical perspectives, dynamics, and characteristics of TPVs, along with their impacts on various meteorological phenomena and processes in the Arctic

and midlatitudes. The next several sections provide background information on the current knowledge of TPVs and their characteristics and dynamics to motivate the subsequent review of their linkages to ACs, PLs, and CAOs.

1.2 TPV history in literature

This section contains a review of literature relevant to TPVs from the 1980s through the introduction of the term TPV by Cavallo and Hakim (2009), with a focus on Arctic tropopause folds, tropopause-based vortices and their linkages to surface cyclones, and an introduction to modern-day TPV literature.

1.2.1 Arctic tropopause folds

The term TPV was first introduced in the literature by Cavallo and Hakim (2009), with the currently recognized definition of a TPV originating from Hakim and Canavan (2005), though the role of tropopause-based vortices in midlatitude and polar meteorology had been recognized long prior to these studies. Notable early research on tropopause-based vortices includes Kleinschmidt (1950a,b, 1951, 1955, 1957), who recognized the important role of these vortices in surface cyclogenesis and their utility in deriving the flow structure at a time when active research was focused on lower-tropospheric baroclinic processes (Hoskins et al. 1985).

A push of notable research on tropopause-based vortices emerged in the 1980s following the “Presidents’ Day Snowstorm” of 18–19 February 1979 (e.g., Bosart 1981; Uccellini et al. 1984), which underwent unforecasted rapid cyclogenesis and produced heavy snowfall nearing 60 cm in portions of the eastern United States. Subsequent studies identified numerous lower and upper tropospheric factors contributing to its unexpected rapid cyclogenesis, such as enhanced antecedent baroclinicity superimposed over the preexisting sea surface temperature gradient off the eastern North American coast (e.g., Bosart 1981), diabatic processes associated with latent heat release (e.g., Bosart and Lin 1984; Boyle

and Bosart 1986; Whitaker et al. 1988), and an antecedent tropopause-based cyclonic vortex and associated tropopause fold upstream of the lower-tropospheric baroclinic zone that preceded the rapid surface cyclogenesis (e.g., Uccellini et al. 1985; Hibbard et al. 1989). This case ushered in a new era of cyclone research relating to rapid cyclogenesis and the relative importance of dry dynamics and diabatic processes (Schultz 2022).

Midlatitude tropopause folding, identified by a downward intrusion of stratospheric air, has been well documented dating to the mid-20th century (Keyser and Shapiro 1986, section 2a). Documentation of Arctic tropopause folds emerged much later, with Shapiro et al. (1984) and subsequently Shapiro et al. (1987) noting that Arctic tropopause folds are characterized by lower altitudes and lower potential temperatures on isobaric surfaces than their midlatitude counterparts. Shapiro et al. (1987) performed a case study of a feature they described as a “polar vortex” that tracked south through Canada into the northern United States. They subsequently introduced a “three-fold” model structure of the tropopause highlighting the Arctic, polar and subtropical jet streams. Retrospective analyses (Walsh et al. 2018; Bray et al. 2021) suggest the Arctic tropopause fold and “polar vortex,” as described in Shapiro et al. (1987), corresponds with what is now known as a TPV, as the feature they termed a “polar vortex” is much smaller spatially than what is now referred to as a stratospheric polar vortex (Vaugh et al. 2017).

1.2.2 Early identification of tropopause-based vortices

Though TPVs only recently entered the meteorological lexicon, they are not a new phenomenon. Factors that may have hindered the identification of tropopause-based vortices included the lack of observations with sufficient horizontal and vertical density in high latitudes, and the lack of sufficient horizontal and vertical resolution to resolve these vortices in numerical models. Hoskins et al. (1985, section 2c) compared two vertical cross sections through an Icelandic low from different data sources (shown in Fig. 1.1), one from the European Centre for Medium-Range Weather Forecasts (ECMWF) analysis on

a 3.75° horizontal grid, which is largely insufficient at resolving subsynoptic scale features such as TPVs (Fig. 1.1a), and one from a high-resolution analysis constructed manually using aircraft and dropwindsonde data (Fig. 1.1b). The high-resolution cross section shows prominent tropopause folds to the northwest and southeast of the low, associated with upper-tropospheric fronts and tropopause-level wind speed maxima, and a depressed tropopause in the center. These tropopause folds are barely identifiable in the accompanying coarse-resolution ECMWF cross section.

The role of vertical resolution in analyzing tropopause-based vortex characteristics was also investigated from an idealized perspective by Thorpe (1986), through comparing cold-core and warm-core vortices, and tropopause-based and surface-based vortices. The cold core, cyclonic tropopause-based vortices (his Fig. 1) exhibit lower tropopause height at the vortex core relative to the surrounding environment, with a positive temperature anomaly in the lower stratosphere and a negative temperature anomaly below the tropopause. Thorpe concluded that improved vertical resolution near the tropopause relative to operational NWP models at the time is important to accurately representing the tropopause and associated synoptic-scale flow, and further suggested some of the idealized vortices in his study may be better described as mesoscale systems.

Nowadays, it is known that TPVs can exhibit exceptionally long lifetimes (Hakim and Canavan 2005). Prior to the introduction of the term TPV, climatologies of cyclonic and anticyclonic upper-tropospheric cutoff troughs identified subjectively (Sanders 1988) and objectively (Bell and Bosart 1989) showed some of these disturbances can exist for over a month. These studies used an isobaric surface, in both cases 500 hPa, which in the context of TPVs does not adequately represent the tropopause. Morgan and Nielsen-Gammon (1998) note that beginning around the late 1980s, the dynamic tropopause surface, defined as a surface of constant Ertel PV (EPV), came into more common usage as the dynamic

tropopause provides an easier alternative to identifying the upper-tropospheric and lower-stratospheric PV distribution on a single surface instead of analyzing PV on multiple isentropic surfaces.

Case studies of the 13 March 1993 “superstorm” (e.g., Bosart et al. 1996) and the 25–26 January 1978 cyclone (e.g., Hakim et al. 1996) using the dynamic tropopause identified numerous long-lived subsynoptic scale tropopause-based vortices, in some cases traced back 10–14 days prior to surface cyclogenesis, which played a major role in the subsequent rapid intensification of the cyclones. The role of these tropopause-based vortices in inducing surface cyclogenesis was further investigated around this time in numerical simulations with idealized vortices (e.g., Takayabu 1991, 1992; Schär and Wernli 1993), motivating the need to further study these features.

1.2.3 Modern analyses of TPVs

Hakim (2000) hypothesized that long-lived tropopause-based disturbances are inherently different from quasi-linear RWs. By partitioning 500-hPa vorticity maxima into weak, moderate, strong and extreme categories, Hakim (2000) showed the strong and extreme categories most closely resemble nonlinear coherent vortices. When collocated with a tropopause potential temperature minimum, the features resemble what we now know as TPVs. A dynamical linkage was also shown between strong and extreme coherent structures and surface cyclones, providing a systematic perspective to the dynamical linkage shown in individual case studies discussed in section 1.2.2.

These tropopause-based cyclonic vortices were later referred to as CTDs by Pyle et al. (2004), which are dynamically the same as TPVs, but as noted by Biernat et al. (2021) lack the high-latitude criteria that TPVs require. Pyle et al. (2004) identified and tracked a long-lived CTD for 17.5 days, during which it aided in intensifying multiple jet streaks. As CTDs and by extension TPVs are cold-core phenomena, as they approach a background

midlatitude jet stream or preexisting jet streak, they have the potential to strengthen the jet via enhancing the PV and tropopause potential temperature gradients.

Hakim and Canavan (2005) used the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis dataset (NCEP/NCAR) (Kalnay and Coauthors 1996) to create the first systematic climatology of tropopause-based cyclonic and anticyclonic vortices, in addition to creating an Arctic subset of these tropopause-based vortices by requiring they spend at least 60% of their lifetime poleward of 65°N. The cyclonic subset became what (Cavallo and Hakim 2009) defined as TPVs. The tropopause-based cyclonic vortices were found to have longer lifetimes than their anticyclonic counterparts, occur more frequently and with a larger amplitude during the winter months than the summer months, and tend to progressively track equatorward with time.

Once TPVs had been defined, research focused on assessing dynamics contributing to TPV intensity changes, where TPV intensity is understood to refer to EPV in the core of the vortex (Cavallo and Hakim 2010). Cavallo and Hakim (2009) suggested that TPV intensity changes in the absence of midlatitude jet interactions are dependent on local physical processes rather than large-scale circulation anomalies. Studies such as Wirth (1995) and Zierl and Wirth (1997) assessed factors that contribute to changes in vortex intensity, with diabatic heating contributing to vortex weakening and radiative cooling contributing to vortex intensification. Cavallo and Hakim (2009) conducted a numerical modeling case study of a long-lived TPV, with their results consistent with Wirth (1995) and Zierl and Wirth (1997) indicating that a TPV intensifies when there is a net dominance of radiative cooling over latent heating, and vice versa for TPV weakening.

1.3 TPV characteristics and dynamics

This section focuses on the structure of TPVs and dynamics controlling their intensity, reviewing theoretical, numerical and observational methods used to diagnose TPV characteristics and dynamics.

1.3.1 TPV characteristics

As TPVs are coherent structures, they exhibit fundamentally different properties than quasi-linear RWs. RWs generally exhibit wavelengths of about 4000 km at 45°N latitude, a westward tilt with height through the troposphere, and a mean eastward phase speed of 12–15 m s⁻¹ (e.g., Hakim 2000; Wirth et al. 2018), whereas TPVs exhibit much smaller radii commonly on the order of 150–600 km (Cavallo and Hakim 2012). Nonlinearity is expressed as the parameter δ_β representing the scaling of nonlinear effects associated with advection (Hakim 2000):

$$\delta_\beta = \frac{U}{BL^2}. \quad (1.1)$$

The equation above represents non-dimensional horizontal scales for length (L), wind (U), and β (B). The vertical vorticity equation is a differential equation with a solution directly proportional to δ_β . Since $\beta \rightarrow 0$ towards the poles, this implies from (1.1) that δ_β becomes large in high latitudes, and thus the right-hand side of the vertical vorticity equation is no longer negligible. For small values of δ_β , this is a linear homogeneous differential equation whose solutions can be described as RWs. Coherent tropopause-based vortices have δ_β of $O(1)$ or greater, and as much as $O(10)$ for strong vortex events. These large values indicate the differential equation is nonlinear and its solutions cannot be described as linear RWs or weakly nonlinear waves (Hakim 2000).

The coherent, nonlinear structure of TPVs allows them to be investigated from the perspective of an idealized theoretical axisymmetric, isolated vortex. This enables for application of the invertibility principle, defined as follows: “If the spatial distribution of IPV is known and the flow satisfies a balance condition, such as geostrophic or gradient wind balance, then it is possible, in principle, to determine the details of that balanced state” (Thorpe 1986). It is understood that isentropic PV (IPV) refers to isentropic maps of EPV. It is shown in Thorpe (1985, 1986) that the invertibility principle can be expressed for an axisymmetric flow in gradient wind balance and hydrostatic balance, and thus allows

for retrieval of the balanced flow, such as perturbation potential temperature, geopotential height and wind, associated with a PV anomaly. Unless otherwise noted, usage of the term “anomaly” from here on refers to an anomaly relative to the background environment.

An early theoretical diagnosis of upper-tropospheric PV anomaly characteristics by Hoskins et al. (1985, section 3) applied the invertibility principle for an idealized, isolated cyclonic (Fig. 1.2a) and anticyclonic (Fig. 1.2b) PV anomaly. A depressed tropopause is apparent for cyclonic PV anomalies, with an elevated tropopause for anticyclonic PV anomalies, and temperature and wind perturbations extend both above and below the PV anomaly in both cases. For a cyclonic vortex, static stability and absolute vorticity are higher at the vortex center relative to the background isentrope. As the sign of absolute vorticity is the same within, above and below the PV anomaly, in order to maintain thermal wind balance, the sign of the static stability anomaly below and above the PV anomaly is the opposite sign of the PV anomaly. For cyclonic PV anomalies, this implies a negative static stability anomaly below and above the tropopause, thus resulting in the characteristic upward bending of isentropes with a cold anomaly below the vortex core. Above the vortex core, isentropes bend downward with a warm anomaly.

The PV inversion approach was further assessed in Thorpe (1986) for various cases of cold core and warm core cyclones and anticyclones, both in the lower and upper troposphere. The cases of a cold core upper cyclone, as shown in their Figs. 1 and 5, exhibit a similar lowering of tropopause height and bending of isentropes as in Hoskins et al. (1985). They additionally noted that when simply imposing an upper cyclone or anticyclone without a corresponding surface anomaly, the surface pressure anomaly resulting from an upper cyclone is greater than the surface pressure anomaly resulting from an upper anticyclone.

The application of PV diagnostics became much more efficient with advances in the horizontal and vertical resolution of numerical models. Using the Advanced Research (ARW) Weather Research and Forecasting (WRF) numerical model (Skamarock et al.

2005), Cavallo and Hakim (2010) created a composite structure of TPVs using 568 samples identified over a 3-year period over northeastern Canada where TPVs most commonly intensify. The resulting composite vortex east-west cross section is shown in Fig. 1.3. Key characteristics of TPVs relative to the background environment are a cold temperature anomaly below the tropopause centered over the vortex core, with a warm anomaly located above the vortex core (Fig. 1.3a); cyclonic flow surrounding the vortex core (Fig. 1.3b); a positive PV anomaly centered on the vortex core above the tropopause and maximized near the background tropopause level (Fig. 1.3c); and a negative relative humidity (RH) anomaly centered above the vortex tropopause indicative of a downward intrusion of dry stratospheric air, and a positive RH anomaly in the troposphere below and east of the vortex, resulting in a locally strengthened vertical RH gradient over the vortex tropopause (Fig. 1.3d).

Despite TPV composite structure being well studied from a numerical modeling framework, obtaining in-situ observations directly within TPVs is a challenge owing to relatively sparse observations in the Arctic Ocean (Sato et al. 2018; Johnson and Wang 2021). Additionally, considerations of a TPV's average forward motion and relatively small radius make it difficult to sample TPVs in twice-daily radiosonde launches. A notable exception is Borg et al. (2020), which used frequent observations atop the Greenland Summit Station to create time-height sections through TPVs passing through the station. Their results were generally consistent with Cavallo and Hakim (2010), albeit with a lack of a positive low-tropospheric RH anomaly, likely due to the lower surface moisture content atop the elevated station at 3.2 km above sea level.

1.3.2 TPV dynamics

Early theoretical studies of upper-tropospheric PV anomalies indicated diabatic processes, specifically radiative cooling and latent heating, affect their intensity. Davis and Emanuel

(1991) and Stoelinga (1996) showed latent heating contributes to the formation of a low-level PV anomaly, and destruction of upper-level PV maximized following the absolute vorticity vector downstream of and above the latent heating source. Accordingly, latent heating contributes to a negative EPV tendency at the tropopause and thus weakening of TPVs, as demonstrated in Wirth (1995) and Hoskins et al. (1985). This also motivated the hypothesis of TPVs weakening in lower latitudes; the reduction in tropospheric static stability below the TPV can result in increased convection and subsequent latent heat release if sufficient moisture is present (Thorpe 1986), which is more likely in lower latitudes where moisture content is higher.

Conversely, cloud-top radiative cooling has been shown to contribute to a positive EPV tendency of an idealized cut-off cyclone, depending on where the cloud-top radiative cooling is relative to the tropopause (Wirth 1995). For the case of an anticyclonic tropopause-based vortex, radiative cooling also contributes to a positive EPV tendency and lowering of the tropopause (Zierl and Wirth 1997). Water vapor would thus be an important contributor to TPV intensity as well, as large vertical gradients in water vapor across the tropopause implies large vertical gradients in radiative cooling (Zierl and Wirth 1997).

As TPVs are associated with a PV maximum, Cavallo and Hakim (2009) showed that TPV intensity changes can be quantified using the EPV tendency equation (Pedlosky 1998):

$$\frac{D\Pi}{Dt} = \frac{D}{Dt} \left(\frac{1}{\rho} \omega_a \cdot \nabla \theta \right) = \frac{\omega_a}{\rho} \cdot \nabla \frac{D\theta}{Dt} + \frac{\nabla \theta}{\rho} \cdot \left(\nabla \times \frac{F}{\rho} \right), \quad (1.2)$$

where EPV is defined as:

$$\Pi = \frac{\omega_a}{\rho} \cdot \nabla \theta. \quad (1.3)$$

In (1.2) and (1.3), ω_a is the three-dimensional absolute vorticity vector, ρ is density, θ is potential temperature, ∇ is the 3D gradient operator, and F is friction. The thermodynamic equation can be decomposed into potential temperature tendency contributions

from radiation (rad), latent heating (lh), PBL processes (pbl), cumulus parameterization (cumulus), and dissipation (mix):

$$\frac{D\theta}{Dt} = \dot{\theta}_{\text{rad}} + \dot{\theta}_{\text{lh}} + \dot{\theta}_{\text{pbl}} + \dot{\theta}_{\text{cumulus}} + \dot{\theta}_{\text{mix}}. \quad (1.4)$$

Assuming that the influence of friction is minimal near the tropopause, then the dominant diabatic contributions near the tropopause are due to radiation and latent heating. Radiation can be partitioned into longwave radiation (LW) and shortwave radiation (SW) (Cavallo and Hakim 2009, 2010, 2013), (1.2) and (1.4) can be combined to produce the following approximation:

$$\frac{D\Pi}{Dt} \simeq \frac{\omega_a}{\rho} \cdot \nabla (\dot{\theta}_{\text{LW}} + \dot{\theta}_{\text{SW}} + \dot{\theta}_{\text{lh}} + \dot{\theta}_{\text{other}}). \quad (1.5)$$

Equation (1.5) shows that the dominant factors affecting TPV intensity change are radiative processes, both LW and SW, and latent heating. Typically, the vertical gradients of radiation and latent heating far outweigh their horizontal gradients, indicating that to a good approximation, vertical EPV tendency structures are proportional to vertical gradients of radiation and latent heating tendencies (Cavallo and Hakim 2009).

Cavallo and Hakim (2009) applied equation (1.5) to a case study of a long-lived TPV, partitioning its life cycle to an intensification and weakening phase. Time-height sections during the intensification phase show a positive EPV tendency due to diabatic processes maximized in the tropopause, largely dominated by radiative cooling maximized near the top of the tropospheric cloud layer, with a smaller negative EPV tendency due to radiative heating maximized below the tropopause. Negative EPV tendency due to latent heating in the tropopause is substantially smaller in magnitude relative to positive EPV tendency due to radiative cooling. Results were less conclusive for the weakening phase, but with stronger contribution to EPV tendency from latent heating in the troposphere than in the intensification phase. These results were consistent with prior hypotheses, highlighting diabatic contributions to TPV intensification are dominated by radiation.

The aforementioned case study motivated a climatology of 568 WRF simulated Arctic TPVs in Cavallo and Hakim (2010), focusing on northeastern Canada where TPV intensification is frequent. Results showed an average tendency for TPVs to diabatically intensify in the Arctic, dominated by radiation forcing (Fig. 1.4a,b). Tropopause-level EPV creation in the vortex core is associated with a stronger vertical radiative heating gradient relative to the surrounding environment, due to anomalously large radiative cooling below the tropopause and anomalously small radiative cooling above the tropopause (Cavallo and Hakim 2010, their fig. 10c,d). Negative EPV tendencies on the tropopause due to latent heating are smaller in magnitude than positive EPV tendency due to radiative cooling (Fig. 1.4c), as the magnitude of latent heating is smaller in the Arctic than midlatitudes owing to lower temperatures and moisture content. Seasonal comparisons show larger negative EPV tendency due to latent heating during summer than winter, as both temperatures and water vapor increase relative to winter, while positive EPV tendency due to radiative cooling is much greater in magnitude and shifted upward into the lower stratosphere during summer relative to winter (Cavallo and Hakim 2010, their fig. 15).

Cavallo and Hakim (2012) further examined the role of LW cooling on TPV intensity, comparing two sets of WRF simulated TPV climatologies with and without radiation. Removing radiation resulted in a 22.3% decrease in vortex amplitude during winter 5 days after genesis, and a 38.0% decrease during summer, but only minimally decreased TPV counts with nearly identical TPV radii, indicating radiation plays an important role in TPV dynamics, as TPVs weaken in the absence of radiative forcing. Another interesting result was that the largest decrease in TPV amplitude without radiation occurred in the summer. As tropospheric moisture is higher in summer than winter, this motivated a hypothesis that moisture plays an important role in modulating TPV intensity.

It is known that low-level clouds are frequent and widespread across the Arctic, with the presence of mixed-phase clouds indicative of at least some presence of supercooled liquid (e.g., Curry et al. 1996; Intrieri et al. 2002; Shupe 2011). Furthermore, cloud-top radiative

Experiment	LW	WV	SW	Clouds
E1	Yes	No	No	No
E2	Yes	Yes	No	No
E3	Yes	Yes	Yes	No
E4	Yes	Yes	No	Yes
E5	Yes	Yes	Yes	Yes

Table 1.1: Summary of experiments from Cavallo and Hakim (2013), denoting inclusion of longwave radiation (LW), water vapor (WV), shortwave radiation (SW), and clouds.

cooling is dominated by its liquid water content, indicating the importance of both liquid-only and mixed-phase clouds to LW radiative cooling (Turner et al. 2018). The question of the role of water vapor in TPV intensity changes motivated an idealized experiment in Cavallo and Hakim (2013) to isolate the contributions of LW radiation, SW radiation, latent heating (LH), water vapor (WV), and clouds. Five experiments were constructed, detailed in Table 1.1, with their resulting vortex amplitudes shown in Fig. 1.5.

Experiment E1, without WV and with LW radiation the only diabatic contribution to EPV tendency, shows only weak TPV intensification of 5% in 5 days with EPV tendency failing to resemble the structure observed in Cavallo and Hakim (2010). Adding WV in experiment E2 results in more substantial intensification of the vortex of 17% in 5 days due to the strong vertical WV gradient, which is characteristic of the vortex core as noted in section 1.3.1. E2 also has a downward intrusion of dry stratospheric air in the TPV core and a depressed tropopause height resulting in an enhanced vertical WV gradient. As WV is a strong absorber of LW radiation and decreases with height, it contributes to an enhanced vertical gradient in radiative cooling and increases tropopause-level EPV generation.

Adding SW radiation in experiment E3 only slightly decreases vortex amplitude compared to E2. As SW radiation is not as strongly absorbed by WV, it contributes to a lower

magnitude in the EPV tendency in comparison to LW radiation. There is a small net negative EPV tendency as SW radiation slightly decreases with height near the tropopause. Experiment E4 removes SW radiation and introduces clouds and their associated LH, resulting in much greater TPV intensification of 34% in 5 days. Despite the role of LH in upper-level EPV destruction, the presence of low-tropospheric clouds with a strong vertical gradient in cloud ice and liquid content results in enhanced LW cooling above the cloud layer. Combining the effects of SW radiation and clouds in experiment E5 reduces TPV intensification from E4 to 19% over 5 days, with less tropopause EPV generation due to lower mean cloud mixing ratios, resulting in less cloud-top radiative cooling. Considering annual variations in Arctic solar radiation, E4 is likely characteristic of a wintertime TPV, with E5 characteristic of a summertime TPV.

1.3.3 TPV preference in high latitudes

The results from Cavallo and Hakim (2009, 2010, 2012, 2013), in conjunction with early theoretical foundations of factors affecting tropospheric PV anomaly intensity (e.g., Thorpe 1986; Stoelinga 1996; Wirth 1995), explain why TPVs are climatologically favored in higher latitudes over the midlatitudes. Zonal mean temperature decreases with increasing latitude, which per the Clausius–Clapeyron equation implies an exponential decrease in moisture and accordingly latent heating with latitude. As TPV intensification occurs when EPV generation from radiative cooling dominates over EPV destruction due to latent heating, TPVs are more likely to intensify, or at least maintain their intensity, in high latitudes where moisture content is much lower than lower latitudes.

As TPVs and their CTD counterparts exit the high latitudes, several factors can contribute to weakening. Higher moisture content and enhanced latent heating result in EPV destruction due to latent heating becoming the dominant diabatic contribution to EPV tendency. Convection associated with latent heating in the troposphere would then result in upper-tropospheric EPV destruction. SW radiation slightly reduces TPV intensification

(Cavallo and Hakim 2013), which likely plays a larger role in favoring TPV intensification during winter as solar radiation is absent in high latitudes while present in lower latitudes. Lastly, TPV intensity changes in the absence of a nearby jet stream are largely a function of local diabatic processes; as TPVs enter the midlatitudes, it becomes more probable that they will interact with the midlatitude jet stream, thus making them susceptible to adverse impacts from horizontal wind shear and deformation induced across the TPV (Cavallo and Hakim 2012). This is further evidenced by TPV lifetimes being significantly longer in summer than winter, when the polar jet stream is weakest (Cavallo and Hakim 2012).

1.4 Tropopause definitions

TPV analyses often require objective feature identification and tracking, which is sensitive to the input data supplied to algorithms that perform these tasks. As TPVs are based on the tropopause, the definition of the tropopause and the surface used to identify TPVs on is of importance. Generally speaking, the tropopause is defined as the separation between the troposphere and stratosphere where atmospheric properties can vary greatly (Wirth 2001; Holton et al. 1995). Common definitions of the tropopause include:

- “Thermal tropopause” (WMO 1957): The lowest level where the vertical temperature gradient $\frac{dT}{dz}$ from that level up to 2 km above it exceeds -2 K km^{-1} ,
- “Potential temperature gradient tropopause” (Tinney et al. 2022): The lowest level where the vertical potential temperature gradient $\frac{d\theta}{dz}$ from that level up to 2 km above it exceeds 10 K km^{-1} ,
- “Dynamic tropopause” (Hoskins et al. 1985; Wirth 2001): Defined along a constant PV surface,
- “Ozone tropopause” (Bethan et al. 1996): Defined by differences in both ozone mixing ratio and vertical gradient of ozone mixing ratio.

Wirth (2001) compared three of these tropopause definitions for upper-tropospheric cyclones and anticyclones, finding the thermal tropopause height differs substantially from the dynamic tropopause height, more so for cyclones than anticyclones, the thermal tropopause is often higher than the dynamic and ozone tropopauses, and the ozone tropopause is often in good agreement with the dynamic tropopause.

The dynamic tropopause is commonly used for TPV identification and tracking, with a common potential vorticity unit (PVU; $1 \text{ PVU} = 10^{-6} \text{K kg}^{-1} \text{m}^2 \text{s}^{-1}$) surface of 2 PVU (Morgan and Nielsen-Gammon 1998; Cavallo and Hakim 2009; Szapiro and Cavallo 2018). Sensitivity testing to the choice of PVU surface shows 19% more cyclones and 24% more anticyclones on the 3.0 PVU than 1.5 PVU surface, with comparisons between the two surfaces showing a match within 500 km for 60% of cyclones and 64% of anticyclones (Hakim and Canavan 2005). For a single case study, Cavallo and Hakim (2009) shows the TPV's minimum potential temperature between the 1.75 and 2.25 PVU surfaces has relatively little sensitivity to the choice of PVU surface, except for a 5 day period when the TPV is at its strongest, as the dynamic tropopause is in close proximity to a surface inversion (Cavallo and Hakim 2009).

1.5 TPV remote dynamics

This section discusses literature relevant to ACs, PLs and CAOs, and their linkages to TPVs, or more broadly upper-tropospheric PV anomalies.

1.5.1 TPV-Arctic cyclone linkages

ACs are commonly defined as synoptic-scale cyclones that either develop in, or move into, the Arctic (Gray et al. 2021). The minimum latitude threshold used to define ACs varies by study, but typically ranges between 50°N and 65°N (e.g., Finocchio et al. 2020; Vessey et al. 2020). These cyclones can be long-lived, on occasion lasting for over 10 days (Simmonds and Rudeva 2012; Yamagami et al. 2017). The seasonal distribution and characteristics of

ACs vary by season, with wintertime ACs commonly entering the Arctic from the midlatitudes via the North Atlantic, and summertime ACs commonly forming near the Eurasian coast and tracking into the central Arctic (e.g., Vessey et al. 2020; Valkonen et al. 2021). While wintertime ACs tend to be more frequent and more intense than summertime ACs (e.g., Zhang et al. 2004; Valkonen et al. 2021), summertime ACs tend to exhibit longer lifetimes than wintertime ACs (e.g., Zhang et al. 2004; Simmonds and Rudeva 2014).

ACs can contribute to sea ice loss, though their impact on sea ice varies by time of year. Finocchio et al. (2022) found that June ACs are associated with less sea ice loss relative to non-AC days, as decreased cloud cover outside of cyclones and lower surface albedo over land results in more absorption of solar radiation and associated warming of the sea ice surface. On the contrary, August ACs are associated with more sea ice loss relative to non-AC days, as strong surface wind promotes faster sea ice drift and upper ocean mixing aiding in melting the sea ice bottom given warmer ocean temperatures. A notable summertime case is the August 2012 “Great Arctic Cyclone”, which was characterized by its extreme intensity, long duration, and contribution to rapid sea ice loss (e.g., Simmonds and Rudeva 2012; Zhang et al. 2013; Tao et al. 2017b,a). Some notable wintertime cases of intense ACs contributing to sea ice loss exist as well, such as a record-strong 932 hPa cyclone near Svalbard in January 2022 (Blanchard-Wrigglesworth et al. 2022).

There are numerous processes unique to the Arctic that are associated with AC intensification. A corridor of strong baroclinicity along the Arctic coastlines of Europe, Asia and North America, referred to as the Arctic frontal zone (AFZ) (e.g., Crawford and Serreze 2016), is associated with locally strong low-tropospheric baroclinicity. While the AFZ does not appear to be associated with a local maximum in AC genesis, its associated local baroclinicity maximum can either increase AC intensification rates or decrease AC decay rates (Crawford and Serreze 2016). Furthermore, years with a stronger AFZ tend to be associated with more intense and more frequent ACs than years with a weaker AFZ (Day et al. 2019).

Given the abundance of TPVs in high latitudes, they also function as a common forcing mechanism for AC intensification. Gray et al. (2021) compared summertime ACs that are matched or unmatched to TPVs based on whether the distance separating them is within twice the Rossby deformation radius during both maximum growth rate and maximum intensity of the AC. A majority of ACs develop without a TPV in proximity, but TPVs play a greater role in the intensification of ACs. Excluding northeast Greenland where AC genesis is most common, genesis regions for matched ACs are most common over the Arctic coastline from eastern Siberia into North America, extending into the North Pole, while unmatched cases genesis regions for unmatched ACs are most common farther east over the Arctic coastline from northern Europe into western Russia. Cyclone track densities are similarly more heavily weighted towards the North American side of the Arctic Ocean for matched ACs, and the Eurasian side of the Arctic Ocean for unmatched ACs.

The vertical structure of intensifying ACs can exhibit pronounced variability depending on time of year and linkages with TPVs. For summertime ACs, Gray et al. (2021) found that matched cases exhibit smaller vertical tilt in vorticity structures and greater lower-tropospheric baroclinicity during the lead up to the peak growth rate relative to unmatched cases. Vessey et al. (2022) compared vertical structure composites of summertime and wintertime ACs prior to, during and after their peak intensity. A pronounced lowering in the composite mean tropopause height exists over summertime ACs during and after their peak intensity relative to wintertime ACs. Additionally, summertime ACs exhibit an axisymmetric cold-core structure during and after their peak intensity located beneath a lower stratospheric warm anomaly relative to the surrounding environment, indicative of the surface cyclone being collocated with a TPV aloft. This vertical alignment of the AC with a TPV aloft is not apparent for the wintertime AC composite. Clancy et al. (2022) did not find this axisymmetric cold-core structure for summertime ACs, though as Vessey et al. (2022) speculates, differences between their results and those of Clancy et al. (2022) may

be attributable to a different compositing methodology and a much larger sample size that is not exclusive to the most intense cyclones.

Croad et al. (2023) approached the question of AC vertical structure differences from a phase space perspective, by partitioning summertime AC vertical structure during development into low-level-dominant (LLD) and upper-level-dominant (ULD) categories. Prior to the AC peak intensity, LLD cases tend to exhibit a warm-core asymmetric structure, while ULD cases exhibit a cold-core asymmetric structure. Additionally, both categories transition to an axisymmetric cold-core structure during and after the AC peak intensity. Their AC track composites for both categories also align with Gray et al. (2021), assuming that ULD cases are more likely to be matched with TPVs than LLD cases, in that LLD cases tend to track closer to the Eurasian coastline while ULD cases tend to exhibit longer lifetimes and track over the open Arctic north of the eastern Siberian and North American Arctic coastline.

AC development can also be associated with multiple surface cyclones and multiple TPVs. The August 2012 “Great Arctic Cyclone” has been of particular interest due to its extreme intensity, anomalous duration, and contribution to rapid sea ice melt (e.g., Simmonds and Rudeva 2012; Zhang et al. 2013). As the AC rapidly deepened it became vertically coupled with the upstream TPV, resulting in a barotropic structure (Tao et al. 2017b,a), consistent with similar findings for matched AC-TPV cases in Gray et al. (2021). Chen et al. (2022) found that this AC, which they named AC2, developed and deepened downstream of a TPV named TPV2, while being positioned to the southwest of a mature cyclone named AC1 in vertical alignment with a TPV named TPV1. While initially, competing thermal advection from both TPVs mitigated the deepening rate of ACs, the fastest deepening rate of AC2 was observed as AC1 merged into AC2 and TPV1 merged into TPV2. Case studies of intense ACs have found numerous cases of AC and TPV mergers, which can prolong AC lifetimes (e.g., Yamagami et al. 2017), as can AC-TPV vertical coupling (Tao et al. 2017a). Furthermore, the potential for multiple TPVs to be located in the

proximity of the surface cyclone, and their competing influences on the deepening rate of the cyclone, can further distinguish rapidly deepening ACs from their midlatitude cyclone counterparts.

ACs also differ from midlatitude cyclones in their predictability, with baroclinic instability being one source of these differences. Capute and Torn (2021) compared predictability of summertime AC and wintertime North Atlantic midlatitude cyclones by selecting North Atlantic cyclones representative of the typical midlatitude cyclone evolution and ACs that spent at least 80% of their lifetime poleward of 70°N. They further filtered cyclones through a time-integrated cyclone amplitude metric to retain the most intense and long-lasting cyclones. In comparing the two cyclone subsets, Atlantic cyclones exhibit greater track predictability while ACs exhibit greater intensity predictability. Furthermore, cyclones with lower predictability are often in environments of greater baroclinic instability than cyclones with greater predictability. One factor that may limit AC predictability is frequent cyclone mergers, which may be common with long duration ACs (Simmonds and Rudeva 2014). From a broader basin-wide Arctic predictability framework, Biernat et al. (2023) also found Arctic low-skill periods tend to be characterized by more frequent and more intense AC activity, with ACs more commonly associated with greater baroclinic instability, than high-skill periods.

1.5.2 TPV-polar low linkages

PLs are broadly defined as mesoscale, short-lived cyclones in high latitudes, often developing over open water near the sea ice edge (e.g., Moreno-Ibáñez et al. 2021). The exact criteria for a cyclone to be classified as a PL varies, though many studies consider PLs to exhibit radii between 100–500 km (e.g., Turner et al. 2003; Stoll et al. 2018). Minimum lifetime criteria for PLs often ranges between 6 h and 24 h, while some studies impose maximum lifetime thresholds such as 40 h (Moreno-Ibáñez et al. 2021). A global PL climatology created by Stoll (2022) using the ERA5 reanalysis (Hersbach et al. 2020) shows

Northern Hemispheric PLs are most frequent between northern Norway and Svalbard over the Nordic Seas, with a slightly smaller maxima near southeastern Greenland and over the Labrador Sea. North Pacific PL activity tends to be less frequent than the North Atlantic on average, although PL frequency maxima also exist over the Sea of Japan, Bering Sea, and the Gulf of Alaska. Northern Hemispheric PLs are most common during the cold season, with no PL activity during the warm season. Based on their PL identification methodology, typical PL vortex radii are around 150 km.

PLs can develop in various environmental conditions and have various forcing mechanisms. PL development is common in environments of baroclinic instability, barotropic instability, or convection associated with marine cold air outbreaks (MCAOs) (e.g., Businger and Reed 1989; Moreno-Ibáñez et al. 2021). PL structure and forcing mechanisms can vary over the course of the life cycle of a PL as well, with Stoll et al. (2020) finding that a March 2008 PL initially developed in a baroclinic environment before reaching a mature convective stage. Some PLs are notable for their hurricane-like structure, a similarity which has warranted numerous comparisons to tropical cyclones (e.g., Emanuel and Rotunno 1989; Kolstad and Bracegirdle 2017). PLs commonly develop a warm core, often due to warm air seclusion, heat fluxes, or adiabatic warming in downdrafts (Moreno-Ibáñez et al. 2021).

In cases of baroclinic development of PLs, upper-tropospheric PV anomalies (hereafter called UPV anomalies, as most studies on UPV anomalies in association with PLs do not explicitly discuss TPVs) can serve as forcing mechanisms for PL development and intensification (e.g., Montgomery and Farrell 1992; Claud et al. 2004; Martin and Moore 2006; Føre et al. 2011; Shimada et al. 2014). Mallet et al. (2013) found that a preceding UPV anomaly is a necessary condition for PL development in weather regimes characterized by lower tropospheric lapse rates in the Nordic Seas. The extent of interaction between PLs and UPV anomalies depends on the Rossby depth, which is a function of horizontal scale of the UPV anomaly and static stability (Moore et al. 1996).

The exact influence an UPV anomaly can have on PL development appears to be case-dependent. In a PL case study by Føre et al. (2011), PV inversion experiments showed a strong sensitivity of PL intensification rate to the UPV anomaly intensity. This differs from Bracegirdle and Gray (2009), which did not find such a strong sensitivity of the PL to the UPV anomaly once the PL had developed. Shimada et al. (2014) examined three case studies where UPV anomalies contributed to favorable environments for PL development. Given that UPV anomalies can induce synoptic-scale thermal advection and circulation at the surface, Shimada et al. (2014) propose that low-tropospheric synoptic-scale cold air advection is needed to cancel out warm air advection induced by the UPV anomaly to prevent a PL in the presence of an UPV anomaly from growing in size into the synoptic scale.

A challenge in forecasting PLs is their poor predictability. Due to the small size of PLs, coarse global NWP models and reanalyses are inadequate at resolving PL intensity, structure, and frequency (e.g., Stoll et al. 2018; Stoll 2022). Horizontal resolution of NWP models and quality of initial conditions can also influence PL forecast skill. McInnes et al. (2011) found that increasing horizontal resolution from 12 km to 4 km, and further to 1 km, significantly improved the modeled structure and intensity of a PL case verified against field campaign and satellite observations. This improvement due to increased horizontal resolution is case dependent, however, as repeating the experiment for another PL case failed to correctly simulate a PL regardless of resolution. McInnes et al. (2011) speculated that increased resolution may add little value if convective processes have less important roles in the PL development, or if the quality of initial conditions used to initialize the model is poor. Stoll et al. (2020) further found that error growth from a 2.5 km resolution AROME-Arctic model is larger than that of the $\tilde{25}$ -km resolution ECMWF HRES model, likely due to sensitivity to convective processes growing upscale in the AROME-Arctic model.

In part due to these challenges, incorporating targeted observations in the vicinity of PLs may not necessarily improve their forecast skill. Aspelien et al. (2011) assimilated IPY-THORPEX field campaign observations of two PLs into an ensemble forecast system. Impacts of these added observations on forecast skill was mixed, as for the first case, assimilating these extra observations improved PL forecast skill at longer lead times, while for the second case, where operational models had poor forecast skill, assimilating these extra observations did not improve PL forecast skill beyond very short lead times.

1.5.3 TPV-cold air outbreak linkages

While TPVs are ubiquitous in high latitudes, the dynamical features associated with them are occasionally transported into the midlatitudes. In constructing a climatology of Northern Hemispheric TPVs transported from high latitudes to the midlatitudes, Biernat et al. (2021) further relaxed the TPV criteria from Szapiro and Cavallo (2018), requiring a vortex to last at least 6 h poleward of 60° latitude with a lifetime of at least 2 days. Their resulting climatology of TPVs transported into midlatitudes (Fig. 1.6c) shows two maxima, over the Hudson Bay and Siberia, indicating these are common pathways for TPVs and their associated cold pools to travel out of high latitudes into the midlatitudes.

CAOs are broadly defined as an incursion of an Arctic airmass into the midlatitudes, resulting in at least several days of anomalously cold temperatures (e.g., Hartig et al. 2023; Biernat et al. 2021; Papritz et al. 2019). Contributors to CAOs range across a wide range of spatial and temporal scales; from a subseasonal-to-seasonal (S2S) scale perspective, the stratospheric polar vortex (SPV) can contribute to synoptic configurations favoring CAOs via sudden stratospheric warming (SSW) events, weak SPV events, or wave reflection events (e.g., Kolstad et al. 2010; Domeisen and Butler 2020). Over North America, Alaskan Ridge and Arctic High weather regimes tend to favor CAOs over the central U.S., with the Alaskan Ridge regime associated with wave reflection events and the Arctic High regime associated with a preceding weak SPV (Lee et al. 2019; Millin et al. 2022).

TPV dynamical linkages to tropospheric cold pools and subsequently CAOs can be tied to the discussion of TPV characteristics and their associated tropospheric cold anomaly in section 1.3.1. While modern research on TPV linkages to CAOs is relatively new given that TPVs were only named in Cavallo and Hakim (2009), the role of tropopause-based vortices in surface cold outbreaks had been recognized long prior. Case studies such as Boyle and Bosart (1983) linked a CAO and accompanying surface anticyclone to a tropopause-based vortex collocated with a thickness minimum, which from a modern perspective is indicative of a TPV collocated with a cold pool. An interesting result from a systematic evaluation of TPV-cold pool linkages by Biernat et al. (2021) is that while 71.7% of cold pools are associated with a TPV, only 24.0% of TPVs are associated with a cold pool. They speculate that this discrepancy may arise from some TPVs being too small or weak to be associated with a notable cold pool, or TPVs that are associated with open trough that fails to meet their criteria for a cold pool.

Given that northern Canada is a common pathway for TPVs to exit the Arctic, North America is favored for TPV, cold pool and CAO linkages. Biernat et al. (2021) found that CAO linkages to cold pools vary by region, with about 40.9%–43.7% of CAOs over the northern U.S. linked to a cold pool, and about 5.3%–16.4% of CAOs over the southern U.S. linked to a cold pool. Combining the linkage between cold pools, CAOs and TPVs into a systematic analysis, they found 32.1%–35.7% of northern U.S. CAOs are linked to cold pools associated with TPVs, while just 4.4%–12.5% of southern U.S. CAOs are linked to cold pools associated with TPVs.

TPV linkages to CAOs also exist in other parts of the North Hemisphere where TPV pathways into the midlatitudes are more common. For CAOs originating in the Fram Strait, Papritz et al. (2019) found 29% of the top 100 CAOs are associated with a TPV, which increases to 40% of the top 40 CAOs. This compares to a 14.5% probability of a TPV within the same region for any randomly chosen day, implying that not only does a statistically

significant linkage exist between CAOs and TPVs, but that the more intense a CAO is, the more likely it is to be associated with a TPV.

TPV intensity is also known to affect CAO magnitude. From a dynamical perspective, the underlying hypothesis is that a stronger TPV is associated with a stronger negative static stability anomaly below the vortex core and a stronger cold anomaly. Lillo et al. (2021) studied a long-tracked intense TPV following a pathway from the Arctic into the northern U.S. in January 2019, performing numerical simulations to intensify the TPV by applying an upper-tropospheric cooling tendency and lower-stratospheric heating tendency, and vice versa to weaken the TPV. Their results show a clear linkage between TPV intensity and CAO magnitude, with a stronger TPV exhibiting a more equatorward track and colder temperatures across the eastern U.S. relative to the control simulation, while a weaker TPV tracks more poleward with warmer temperatures across the eastern U.S. (their Fig. 14).

Backward air parcel trajectories can also help in establishing a dynamical link between TPVs and CAOs. Papritz et al. (2019) examined a January 1981 CAO in northern Europe associated with a TPV intensifying over the Arctic prior to entering the midlatitudes, forming a low-tropospheric cold pool which similarly strengthened in the process. A majority of backward air parcel trajectories from the peak of the CAO originated in the vicinity of the TPV 5 days prior. Parcels tended to remain collocated with the TPV, were concentrated near the lower tropopause, and slowly descended with time. The previously noted January 2019 North American CAO case study by Lillo et al. (2021) showed consistent results, with backwards parcel trajectories from the TPV indicative of parcels constrained within the TPV circulation as the TPV exited the high latitudes (Fig. 1.7).

1.6 Outline and Objectives

The results of this dissertation are split into three chapters, with background discussion and associated hypotheses discussed below for each chapter.

1.6.1 Systematic examination of rapidly deepening ACs

An important aspect of comparing systematic studies is understanding differences in their methodologies. Vessey et al. (2022) used the 100 most intense ACs per season for their composites, Gray et al. (2021) used the 200 most intense matched and unmatched ACs, and Croad et al. (2023) retained all ACs with a lifetime greater than 24 h and with a track longer than 1,000 km. Furthermore, spatial compositing methodologies differ as well, as Gray et al. (2021) rotated composites to the cyclone motion vector, while Croad et al. (2023) created a phase space metric combining the low- to mid- tropospheric baroclinic asymmetry and a thermal Rossby number term dominated by the vertical gradient of relative vorticity.

While all of these studies analyzed the maximum growth rate of ACs in their composites, it is speculated that there is a difference between compositing the maximum growth rate of the most intense ACs, and the maximum growth rate of the most rapidly deepening ACs. For example, it is possible some intense ACs only gradually intensify over a long duration, while ACs that are moderately intense but not among the top percentile of most intense ACs exhibit shorter but more rapid intensification episodes. Given that a substantial overlap between the most intense ACs and most rapidly deepening ACs is expected, it is hypothesized that the overall characteristics of rapidly deepening ACs (henceforth named RD ACs) are similar to those of the peak deepening rate of the most intense ACs.

Some early studies of midlatitude “bomb” cyclones, such as Roebber (1984), assumed that these cases are produced by mechanisms “distinct from ordinary baroclinicity.” This hypothesis was likely motivated in part by NWP limitations in accurately simulating rapid cyclogenesis at the time (Uccellini 1990). With time, the consensus view came to be that rapid cyclogenesis is a combination of forcing mechanisms associated with ordinary cyclogenesis that act constructively and exhibit sufficiently large magnitudes to result in extreme intensification rates. As Uccellini (1990) stated, “the individual dynamic and diabatic processes should be looked upon as necessary for rapid cyclogenesis, but not sufficient to produce these storm systems when acting alone.” It is hypothesized that a similar conclusion

can be drawn for ACs and processes unique to AC intensification. Accordingly, the aim of this study is to create a climatology of RD ACs to compare against existing climatologies of the peak deepening rate of the most intense ACs, specifically focusing on the evolution of the AC characteristics throughout the RD episode and primarily assessed through lower-tropospheric baroclinic instability and proximity to nearby TPVs. This climatology of RD ACs is compared across seasons, locations, linkages with TPVs, and finally compared with ordinary intensifying ACs.

It is assumed that the overall climatology of RD ACs should be similar, but not identical, to a climatology of the greatest deepening rate of the most intense ACs. The hypotheses of this experiment are as follows:

- A1: Summertime RD cases are associated with stronger forcing from a nearby TPV, longer lifetimes, and closer proximity of the nearest TPV during the RD episode relative to wintertime cases, similarly to climatologies of intense ACs from Vessey et al. (2022).
- A2: RD cases along the poleward extension of the North Atlantic climatological baroclinic zone exhibit longer lifetimes before the RD episode, have a greater relative importance of lower-tropospheric baroclinic forcing, are more favored during the winter, and are farther from the nearest TPV during the rapid deepening episode relative to Arctic cases.
- A3: Cases that begin RD from the onset of when the cyclone is first trackable are more favored to be in the Arctic basin and are closer to the nearest TPV during their onset, relative to cases that exist at least 24 h prior to the onset of RD.
- A4: Processes associated with RD ACs are not inherently different than ordinary deepening ACs, but are associated with larger magnitudes of differential cyclonic vorticity advection and lower-tropospheric baroclinic instability, with constructive interference resulting in greater cyclone intensification rates, and are more likely to

be associated with a nearby TPV during their peak growth rate than ordinary deepening ACs.

1.6.2 Quantifying TPV intensity linkage to cyclogenesis

After systematic statistical linkages between TPVs and RD ACs have been examined, the relationship between TPV intensity and cyclone intensification rates will be examined more closely with the use of numerical simulations. The 2012 “Great Arctic Cyclone” is a candidate case for further evaluation given its association with two TPVs (Chen et al. 2022). While numerous studies have examined the linkage between this AC and the associated TPVs (e.g., Simmonds and Rudeva 2012; Tao et al. 2017a; Chen et al. 2022), this project aims to quantify the role of the intensities of both TPVs in the AC evolution.

In addition to the AC case, the role of TPV intensity in the evolution of an intense PL case will be examined as well. Previous literature on PL linkages to UPV anomalies, for the most part, have not explicitly looked at TPVs, and it is unclear if some of the UPV anomalies in these studies can be considered as TPVs or CTDs. Furthermore, while recent literature has evaluated TPV linkages with ACs, there is little recent literature from a TPV framework linking TPVs to PL evolution.

For both AC and PL cases, the following hypotheses will be tested using numerical simulations:

- B1: A more intense TPV results in a greater peak deepening rate, stronger peak intensity, and an earlier peak intensity of the surface cyclone relative to the control simulation, with the opposite true for a weaker TPV.
- B2: For the AC case, having two nearby TPVs being more intense result in a greater peak intensity of the AC relative to the control simulation or intensifying each TPV individually.

- B3: For the PL case, a weaker TPV results in a delayed occlusion of the surface cyclone and transition to an axisymmetric warm core, and a weaker cyclone overall, relative to the control simulation.

1.6.3 February 2021 CAO linkage to TPVs

A major CAO occurred in February 2021 over North America, notable for its extremity in terms of magnitude and duration (Albers et al. 2022; Millin and Furtado 2022). Large swaths of the contiguous United States (U.S.), in particular the southern Great Plains region, experienced anomalous cold, with temperatures below -17.8°C (0°F) recorded as far south as central Texas. Portions of Texas also recorded their longest consecutive streak of temperatures at or below 0°C on record (Bolinger et al. 2022). This combination of record cold temperatures, an exceptionally long duration of such cold temperatures, and multiple high impact snow and ice storms contributed to this event becoming the most expensive U.S. winter storm on record at the time (Bolinger et al. 2022), as well as a deadly event in large part due to mass failures of power generation units in Texas (Albers et al. 2022).

Several recent studies assessed the linkages between the February 2021 CAO and the SPV evolution. From an S2S perspective, conflicting evidence exists whether a preceding SSW contributed to the CAO, with Davis et al. (2022) assessing a lack of surface impacts linked to the SSW, while Zhang et al. (2022) identify a linkage between the SPV structure and location and subsequent evolution of the beginning of the CAO over Canada. Stronger evidence for synoptic-scale linkages to the CAO exist in shorter timescales, with improved predictability of preceding wave breaking events in the Arctic and Greenland associated with improved extended range predictability of the CAO (Millin and Furtado 2022). While these studies address the large-scale contributors to this CAO, there remain unanswered questions regarding the role of smaller-scale influences impacting the magnitude of the CAO in shorter timescales; this is known to be related to Tropopause Polar Vortices (TPVs), which will be the focus of this study.

The February 2021 CAO case was unique in that it featured two TPVs rotating about each other and subsequently merging over southern Canada. From an idealized framework, cyclonic vortices in close proximity tend to repel each other (Hakim et al. 2002), although in the real atmosphere TPV mergers are known to occur (e.g., Szapiro and Cavallo 2018; Bray et al. 2021). The wind field associated with PV anomalies extends outward horizontally and vertically, thus PV anomalies in close proximity advect each other. This is analogous to a similar scenario in tropical cyclones, commonly referred to as binary interactions (e.g., Yang et al. 2008). Given that the steering flow each TPV imparts on the other is associated with the strength of its circulation, which is a function of its amplitude, it is expected that modifying the intensity of both TPVs will affect the merger rate of the TPVs, the CAO magnitude via the strength of the cold pool, and the surrounding synoptic waveguide.

The hypotheses for this experiment are as follows:

- C1: Stronger TPVs are expected to merge at a faster rate than weaker TPVs.
- C2: Stronger TPVs are expected to lead to a stronger cold outbreak in the southern Great Plains than weaker TPVs, as shown in previous TPV-CAO linkage studies (Papritz et al. 2019; Biernat et al. 2021; Lillo et al. 2021).
- C3: The presence of multiple TPVs and their associated merger prolonged the duration of the CAO in the southern Great Plains, as a faster merger would be associated with the merged TPV propagating downstream earlier, and accordingly a shorter duration of the CAO in the southern Great Plains, than a slower merger where the spatial extent of the area encompassed by both TPVs is larger than that of a single TPV.

1.6.4 Dissertation outline

The outline of the remainder of this dissertation is as follows. Chapter 2 reviews the data sources and methods used for subsequent chapters. Chapter 3 consists of a systematic

evaluation of rapidly intensifying ACs and their linkages to TPVs. Chapter 4 presents two case studies of TPV linkages to an intense AC and intense PL. Chapter 5 presents a case study of an extreme CAO and its linkages to two merging TPVs. Finally, chapter 6 presents the main conclusions of this dissertation and suggestions for future research.

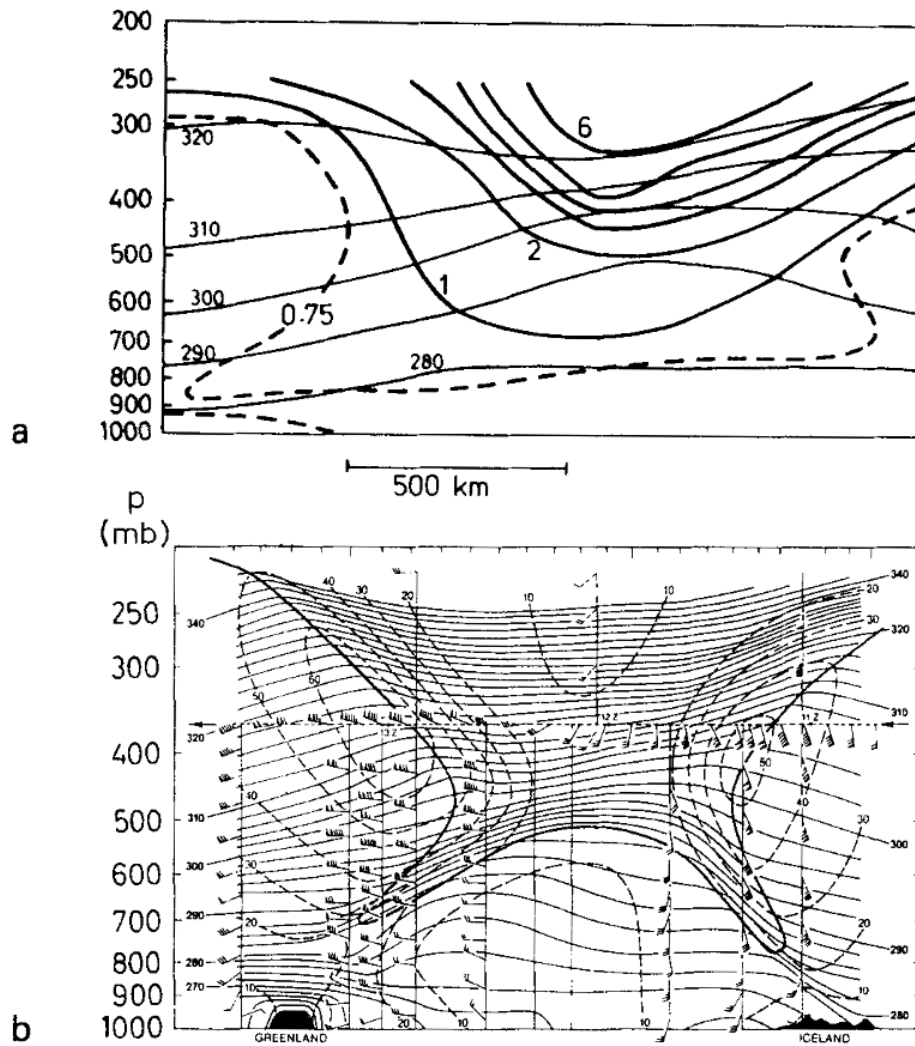


Figure 1.1: Northwest-southeast oriented cross sections through an Icelandic low on 12 April 1983 using two datasets for approximately the same analysis time. (a) ECMWF analysis, with isentropes in thin black contours every 10 K and PV in thick black contours. The thin lines indicate isentropes at 10 K intervals and the thick lines PV at 1 unit intervals. The 0.75 PVU unit contour ($1 \text{ PVU} = 1 \times 10^{-6} \text{ K kg}^{-1} \text{ s}^{-1} \text{ m}^2$) is shown by a heavy dashed line. (b) In-situ observations from aircraft and dropwindsondes, with isentropes in thin black contours every 2 K, isotachs in dashed lines every 10 m s^{-1} , and estimated tropopause in solid black line. Figure reproduced from Hoskins et al. (1985), their Fig. 9.

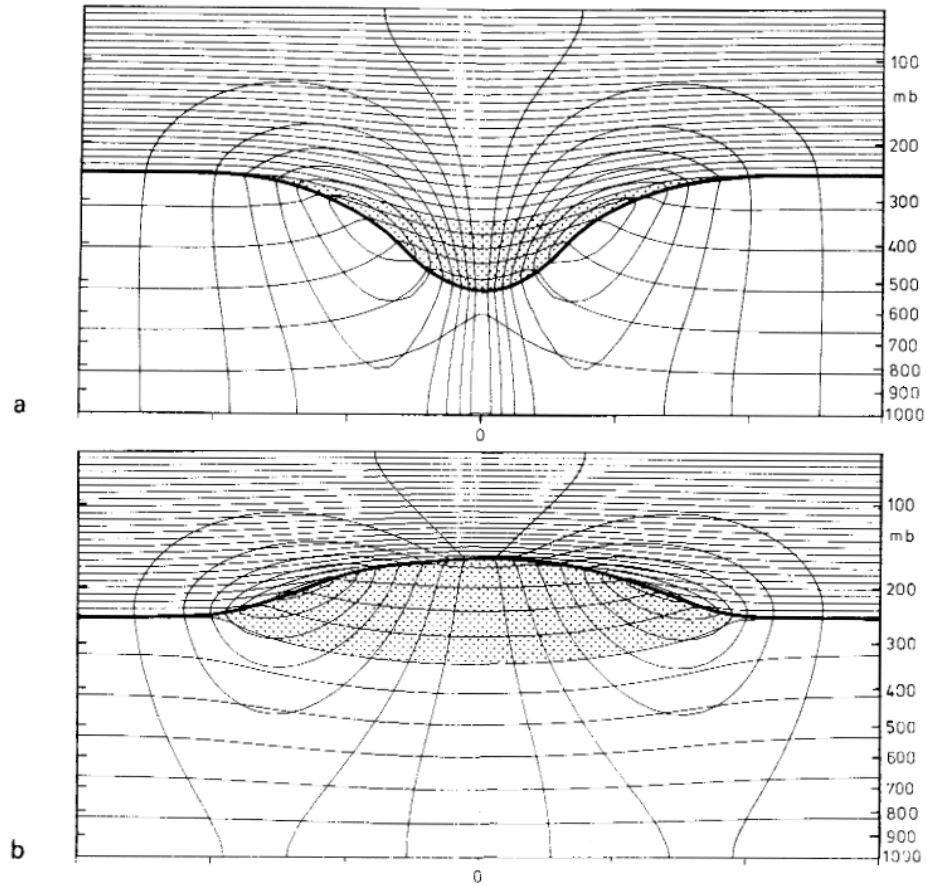


Figure 1.2: Cross section through an isolated upper-tropospheric (a) cyclonic and (b) anticyclonic PV anomaly. The location of the anomaly is denoted by stippled regions. The thick black contour represents the tropopause, and the two sets of thin contours denote potential temperature every 5 K and transverse velocity every 3 m s^{-1} . Figure reproduced from Hoskins et al. (1985), their Fig. 15.

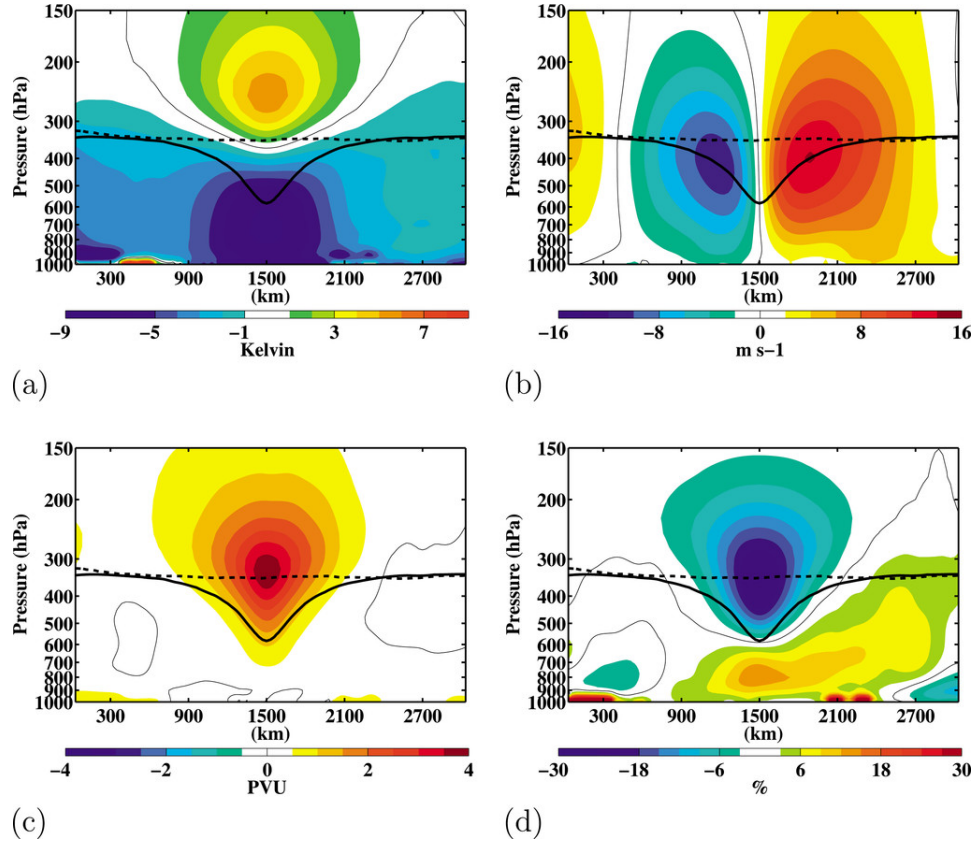


Figure 1.3: Composite TPV west-east cross-vortex sections of the anomalous (a) temperature (K), (b) v-wind component (m s^{-1}), (c) Ertel potential vorticity (PVU), and (d) relative humidity (%). The thick, solid (dashed) black contour is the composite (background) tropopause. The 0 contour is denoted by the thin, solid contour. Figure and caption reproduced from Cavallo and Hakim (2010), their Fig. 9.

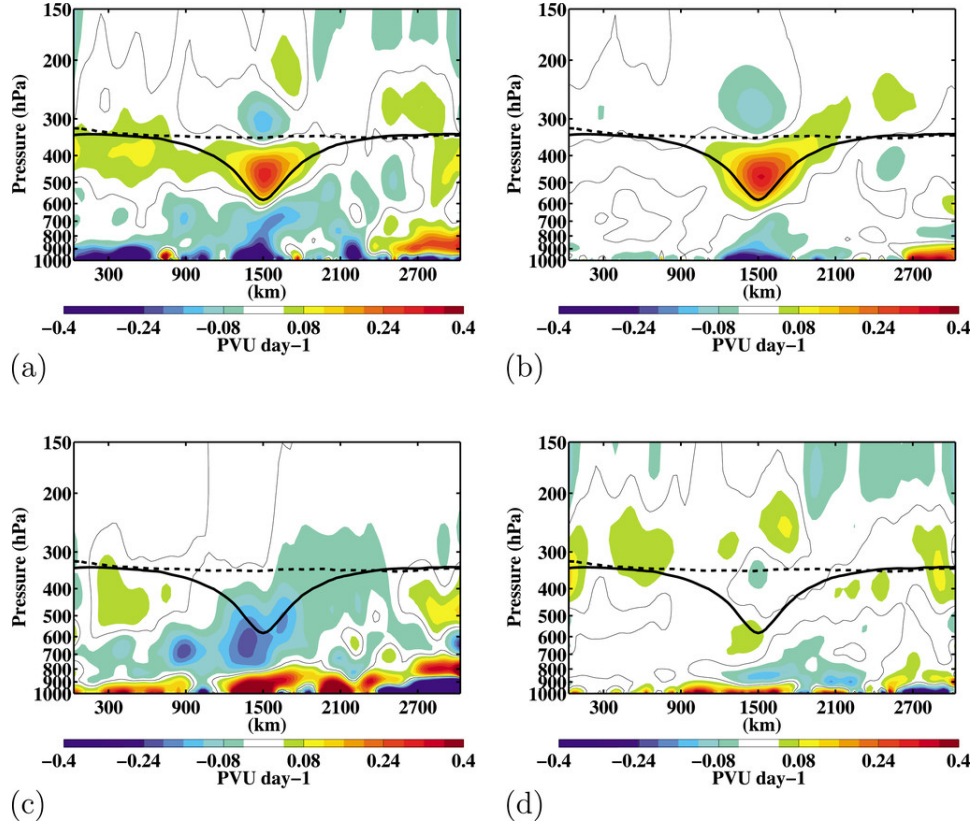


Figure 1.4: Composite TPV west-east cross-vortex sections of the anomalous EPV tendency due to (a) all diabatic effects, (b) radiation, (c) latent heating, and (d) all diabatic effects except for radiation and latent heating. Units are PVU day^{-1} and the contour interval is $0.04 \text{ PVU day}^{-1}$. The thick, solid (dashed) black contour is the composite (background) tropopause. The 0 contour is denoted by the thin, solid contour. Figure and caption reproduced from Cavallo and Hakim (2010), their Fig. 9.

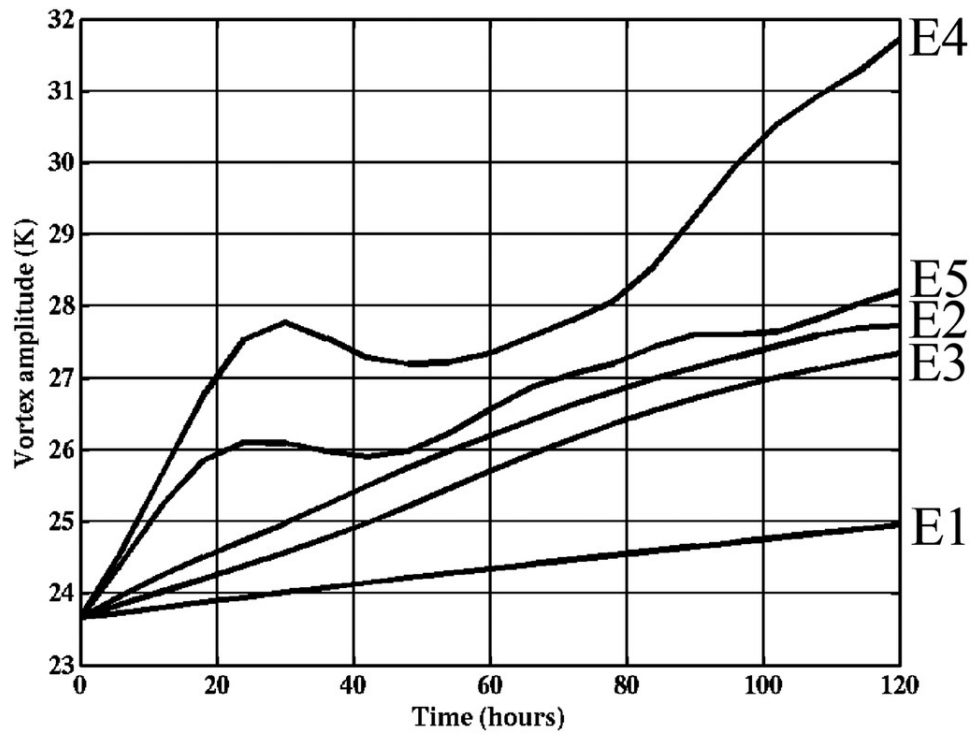


Figure 1.5: Time series plots of vortex amplitude for experiments E1–E5 as labeled in the right margin. Amplitudes are computed on the 2-PVU surface. Figure reproduced from Cavallo and Hakim (2013), their Fig. 3.

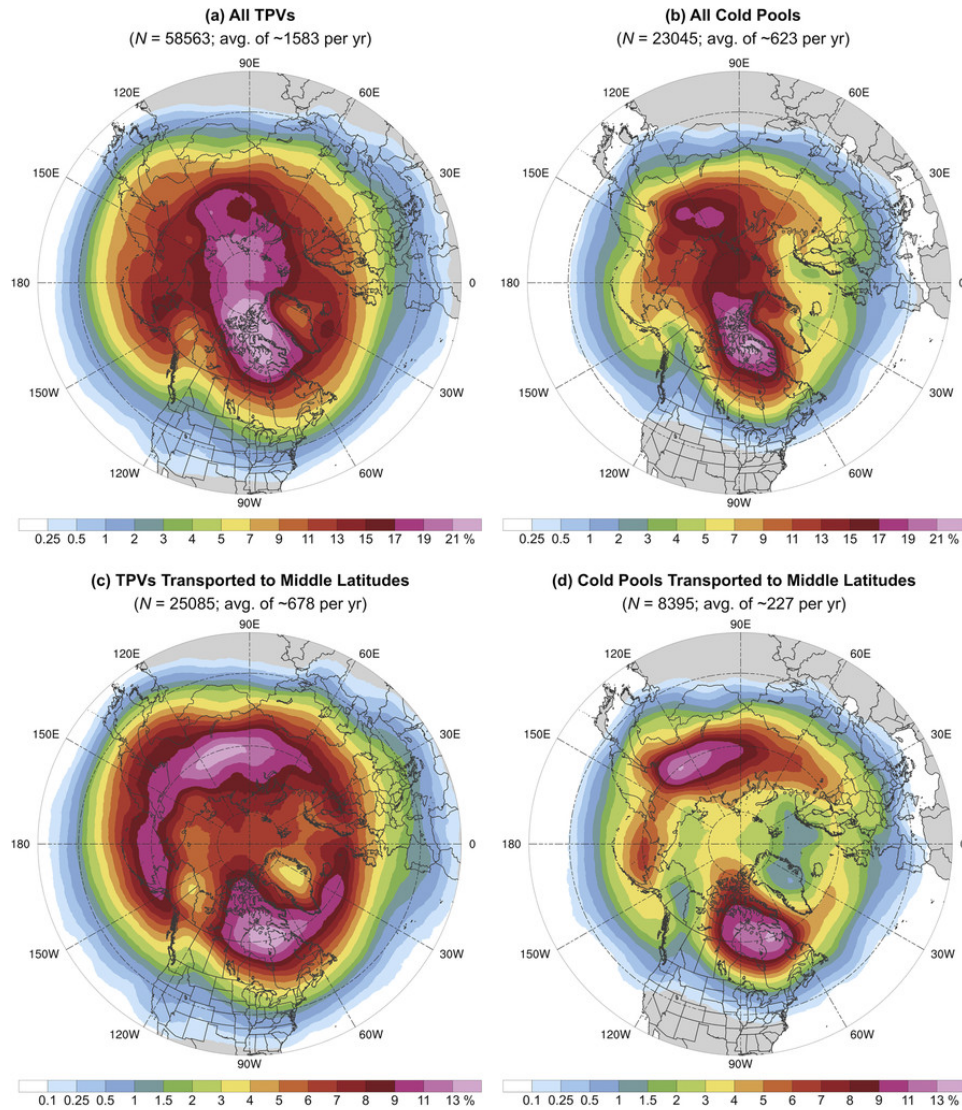


Figure 1.6: Occurrence frequency (%) of (a) all TPVs, (b) all cold pools, (c) TPVs transported to middle latitudes (equatorward of 60°N), and (d) cold pools transported to middle latitudes (equatorward of 60°N). Figure reproduced from Biernat et al. (2021), their Fig. 7.

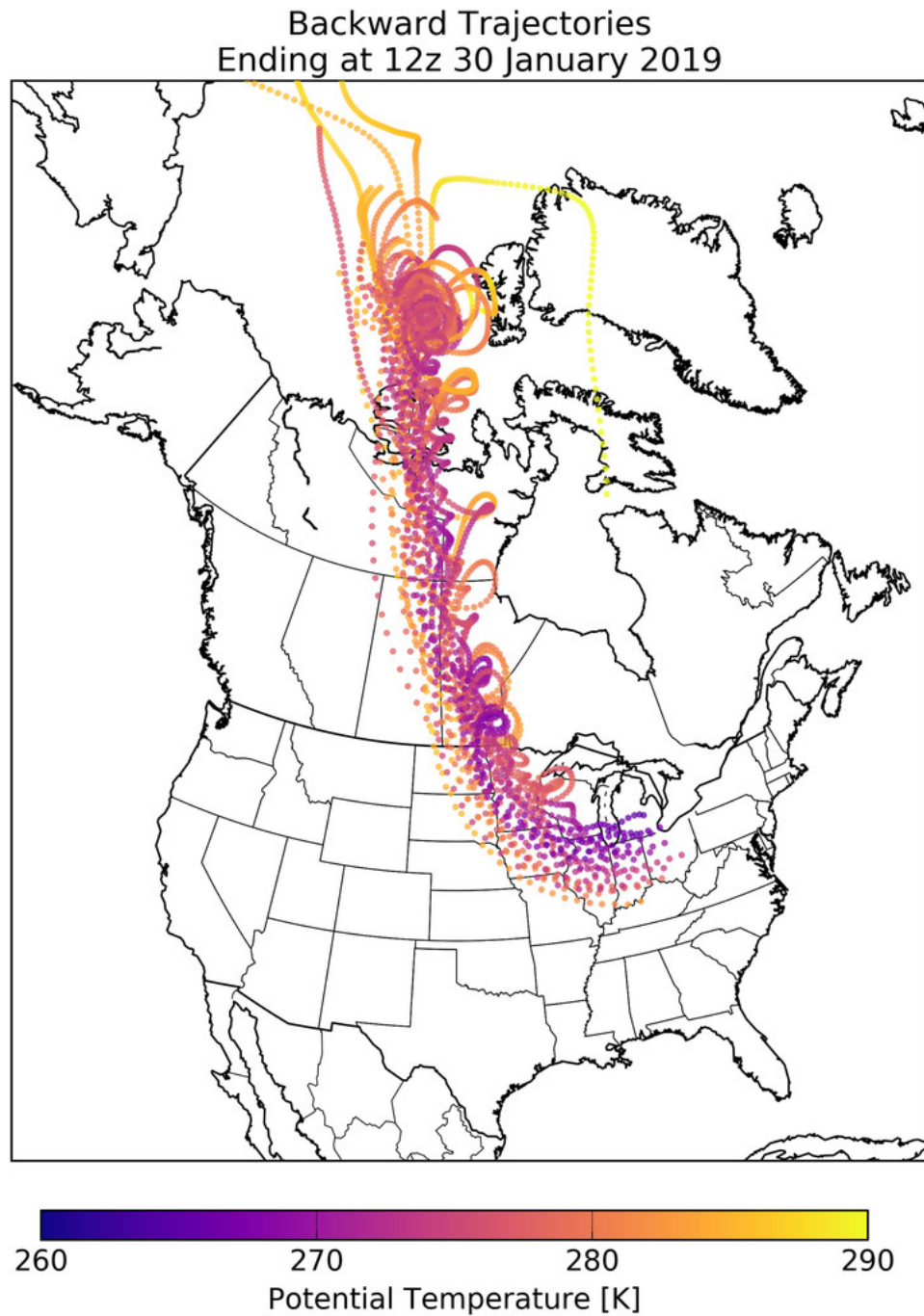


Figure 1.7: Backward trajectories initiated at 1200 UTC 30 Jan 2019, with dots plotted at 1-h intervals and colors representative of the parcel potential temperature. Figure reproduced from Lillo et al. (2021), their Fig. 5.

Chapter 2

Data and methods

This chapter discusses the data sources and methods used for the next several results chapters. Section 2.1 discusses the data sources used for all results chapters, AC and TPV track sources and limitations, and various relevant calculations. Section 2.2 discusses the methodology used to identify RD ACs, generate AC-centered composites, and match ACs to nearby TPVs. Finally, section 2.3 discusses the NWP model setup and experiment design for case study simulations designed to study the linkages between TPVs and an AC, PL, and CAO.

2.1 Data sources

All gridded data analyses in this dissertation use the ERA5 reanalysis (Hersbach et al. 2020). ERA5 has a native horizontal resolution of 31 km, with gridded data available at 0.25° horizontal grid spacing. Advantages to using ERA5 include its substantial improvement in horizontal resolution over previous ERA reanalyses, hourly temporal resolution, and an initial condition (IC) uncertainty ensemble. Limitations include larger uncertainty in the analysis quality in high latitudes and in earlier years of the reanalysis.

TPV tracks and TPV object shapes are obtained using the TPVTrack algorithm (Szapiro and Cavallo 2018), discussed in more detail in section 2.1.4, with the algorithm applied to ERA5 6-hourly data. Surface cyclone tracks are obtained from the Crawford et al. (2021) cyclone tracking and identification algorithm applied to the ERA5 data from 1979–2022. Surface cyclone tracks based on ERA-Interim were used as a similar dataset using ERA5 exhibited numerous spurious topographical features that interfered with the reliability of the cyclone tracks in areas near mountainous terrain.

2.1.1 Calculations

For applications involving upper-tropospheric synoptic forcing for ascent, the quasi-geostrophic (QG) omega equation is used. Several forms of the QG omega equation exist, each with its advantages and limitations. To isolate the contribution of differential geostrophic absolute vorticity advection by the geostrophic wind to forcing for ascent, the right-hand side of the traditional QG omega equation is used, defined by Bluestein (1992, his Eq. 5.6.11) as follows,

$$\left(\nabla_p^2 + \frac{f_0^2}{\sigma} \frac{\partial^2}{\partial p^2} \right) \omega = \underbrace{-\frac{f_0^2}{\sigma} \frac{\partial}{\partial p} \left[-\vec{V}_g \cdot \nabla_p (\zeta_g + f) \right]}_{\text{Term A}} - \underbrace{\frac{R_d}{\sigma p} \nabla_p^2 \left(-\vec{V}_g \cdot \nabla_p T \right)}_{\text{Term B}}, \quad (2.1)$$

where f_0 is the Coriolis parameter, σ is static stability, p is pressure, ω is vertical velocity, $\vec{V}_g$ is the geostrophic wind vector, ∇_p is the gradient operator on an isobaric surface, ∇_p^2 is the Laplacian operator on an isobaric surface, ζ_g is geostrophic relative vorticity, R_d is the gas constant for dry air, f is planetary vorticity, and T is temperature. The right-hand side of equation 2.1 can be partitioned into term A, denoting differential geostrophic absolute vorticity advection by the geostrophic wind, and term B, denoting the Laplacian of temperature advection by the geostrophic wind.

These QG omega terms are typically computed in midlatitudes, as coarse spatial data is needed to smooth out high-resolution noise. For polar applications, traditional grid point spatial smoothing functions are difficult to apply to standard latitude-longitude grids where grid points converge near the pole. To account for this issue near the pole, 500-km area averaging weighted by the area of each grid point within that radius is performed for each grid point.

For applications involving low-level baroclinicity, EGR is commonly used as a metric to quantify baroclinic instability. EGR is defined by Hoskins (1990) as follows,

$$\text{EGR} = 0.31 \frac{f}{N} \left| \frac{du}{dz} \right|, \quad (2.2)$$

where f is the Coriolis parameter, N is the Brunt-Vaisala frequency, and $\frac{du}{dz}$ is the vertical gradient of the zonal wind. For this dissertation, EGR is calculated on the 850-hPa isobaric surface. Due to mountainous terrain affecting EGR distributions, EGR is masked out for all locations above 500 m elevation above sea level.

Many meteorological calculations for this dissertation, such as vorticity and the Laplacian operator, are calculated using the MetPy (May et al. 2022) Python package. Calculations involving Eady growth rate (EGR) use the NCAR command language (NCL). For the 2021 CAO case, air parcel trajectories are computed using the the HYSPLIT program (Stein et al. 2015) applied to 3-h Global Forecast System (GFS) analyses with 0.25° horizontal grid spacing.

2.1.2 Ensemble sensitivity analysis

Some experiments in this dissertation are intended to specifically quantify the sensitivity of the feature of interest to relevant upstream features. A useful method to do so is ensemble sensitivity analysis (ESA) (Ansell and Hakim 2007; Torn and Hakim 2008), which is essentially a lagged linear regression between a dependent variable and an independent variable. Torn and Hakim (2008) define ESA as follows,

$$\frac{\partial J}{\partial x} = \frac{\text{cov}(J, x)}{\text{var}(x)}, \quad (2.3)$$

where J is a response function and x is a sensitivity variable. More simply, J can be interpreted as a variable of interest at a specified domain and time, for example the leading mode of variability of the mean sea level pressure (MSLP) field within a domain surrounding a cyclone of interest at 96 h forecast lead time. Continuing this example, x would be a variable that is expected to have a dynamical linkage to J , for example 500-hPa geopotential height at 24-h forecast lead time. Locations where $\frac{\partial J}{\partial x}$ is significantly different than zero indicate that variability of the MSLP field within the selected domain at 96 h forecast

lead time is sensitive to 500-hPa geopotential height variability in those locations at 24-h forecast lead time.

ESA has been widely used over the last decade and a half, with many applications including, but not limited to, operational tropical cyclone forecasting, convective forecasting, and assessing sources of downstream forecast variability (e.g., Torn 2014; Berman and Torn 2019; Coleman and Ancell 2020; Ancell and Coleman 2022). For this dissertation, ESA is applied for some experiments that make use of ensembles, which is discussed in more detail in their relevant sections in this chapter.

2.1.3 Cyclone tracking and identification overview

While no cyclone identification and tracking algorithms were run as part of this dissertation, given that objective cyclone tracks are used, it is nonetheless important to review the basic principles and limitations of cyclone identification and tracking. In general, there is no single universally accepted method of how to identify an extratropical cyclone (Neu et al. 2013), in contrast to universally accepted cyclone identification and tracking methods for tropical cyclones (Knapp et al. 2010). Major limitations for extratropical cyclone identification and tracking include large vertical tilts, variable cyclone asymmetry and vertical depth, and cyclone merger and splits (Neu et al. 2013).

Some common variables used to identify extratropical cyclones include MSLP (e.g., Wernli and Schwerz 2006), 850-hPa area-averaged cyclonic vorticity (e.g., Hodges 1994; Sinclair 1997), or a combination of MSLP and vorticity (e.g., Hewson and Tittley 2010). Given the variable vertical tilt and depth of extratropical cyclones, in addition to different thresholds used to identify a cyclone, different methods can yield different cyclogenesis and cyclolysis times.

Cyclone tracking is sensitive to numerous factors such as the temporal resolution of the input dataset and the method of matching nearby cyclones between time steps (Neu et al. 2013). Common methods to match cyclones into temporally consistent tracks include

extrapolation based on the ambient steering flow (e.g., Hewson and Tittley 2010), or an extrapolation of the cyclone motion vector and climatological cyclone motion vectors (e.g., Sinclair 1997).

A comparison of various extratropical cyclone identification and tracking algorithms as part of the Intercomparison of Mid Latitude Storm Diagnostics (IMILSAT) (Neu et al. 2013) found major differences were common for metrics including the lifespans of short-lived cyclones, spatial smoothing of input data which is intended to minimize the impact of mesoscale features but at the cost of potentially combining distinct surface cyclones into a single cyclone, and cyclone splits and mergers. While this dissertation uses a single cyclone track dataset for consistency, results of any study that uses an objective cyclone identification and tracking algorithm are inherently sensitive to the algorithm methodology, especially when relating to cyclone genesis, lysis, and duration.

2.1.4 TPV identification and tracking

TPV tracks are obtained using TPVTrack (Szapiro and Cavallo 2018), which utilizes a watershed basin segmentation method to identify TPV shapes, and matches tracks based on temporal overlap of TPV basins. As with extratropical cyclone identification and tracking, this method is sensitive to the spatial and temporal resolution of the input data, which Szapiro and Cavallo (2018) reviews in their Fig. 6. A unique challenge for TPV tracking, which to an extent is also applicable for ACs, is the frequent merging and splitting of TPVs. While no consensus exists for how to best account for mergers and splits, which affects applications such as TPV tracks, TPV genesis, and TPV lysis locations, this is of less importance for applications such as climatological TPV location density and identifying the nearest TPV to a feature of interest, that do not depend on where a TPV track is considered to have started or ended.

The spatial climatology of TPVs is also sensitive to the criteria used to define a TPV, as discussed in section 1.1. Figure 2.1 compares TPV occurrences between the original

criteria from Cavallo and Hakim (2009) and the relaxed criteria from Szapiro and Cavallo (2018) for the northern and southern hemispheres. While the spatial distribution of TPVs poleward of 65° latitude is similar between both methods, the methodology of Szapiro and Cavallo (2018) results in much larger TPV counts extending into the midlatitudes. This motivates using the latter methodology for TPV applications outside of the Arctic and is the basis of TPV tracks used in this dissertation.

2.2 Rapidly deepening ACs

2.2.1 Review of rapid deepening criteria

To assess the characteristics of RD ACs, the first step is to establish RD criteria that is applicable in the Arctic. According to Sanders and Gyakum (1980), Tor Bergeron was reputed to have initially defined a midlatitude explosively deepening cyclone as a cyclone whose central minimum sea level pressure decreases at a rate of at least 1 hPa h⁻¹ for 24 hours. This motivated the original definition of a “bomb” by Sanders and Gyakum (1980),

$$\text{bergeron} = \frac{\Delta P}{24\text{h}} \frac{\sin 60^\circ}{\sin \phi}, \quad (2.4)$$

where ΔP denotes the 24-h change in cyclone central minimum sea level pressure, and ϕ denotes the mean latitude of the cyclone center. This builds upon the original definition by Bergeron by normalizing the cyclone deepening rate to a reference latitude of 60°N, approximately the latitude of Bergen, Norway (Sanders and Gyakum 1980). This normalization accounts for the geostrophic wind variation by latitude given the same pressure gradient (Gyakum et al. 1989).

This definition was slightly adapted in subsequent analyses of explosively deepening midlatitude cyclones for different applications. The reference latitude was modified to 42.5°N by Roebber (1984) and 45°N by Gyakum et al. (1989), as these represent the approximate latitude of the peak observed midlatitude cyclone deepening rates. Yoshida and

Asuma (2004) maintained the 60°N reference latitude but adjusted the temporal threshold to 12 h instead of 24 h to refine focus over the portion of the greatest deepening rate of the cyclone.

A more substantial adaptation of the Sanders and Gyakum (1980) “bomb” criteria was done by Lim and Simmonds (2002), which they called a “relative bomb” and defined as follows,

$$\text{relative bomb} = \frac{\Delta P_r \sin 60^\circ}{24\text{h} \sin \phi}, \quad (2.5)$$

where ΔP_r denotes the change in central minimum MSLP anomaly with respect to the climatological MSLP. A further modified definition was used by Allen et al. (2010), requiring an EC to meet the definition of both (1) and (2).

While the latitude normalization method works for identifying RD midlatitude cyclones, this approach has limitations in high latitudes. Figure 2.2 shows the Bergeron criteria as a function of latitude. The criteria continue to increase as a function of latitude up to the North Pole, where the threshold is approximately 27.5 hPa per 24 h.

Figure 2.3 shows the cyclone track densities of the portions of a cyclone’s life cycle that exceed at least 1 Bergeron. It is evident that these criteria capture very few cyclones in the Arctic, but not for lack of intense cyclones or cyclones that quickly deepen. This indicates that Bergeron criteria substantially exceeds the observed peak cyclone deepening rates, motivating the need for different criteria that capture the top observed cyclone deepening rates in the Arctic.

2.2.2 Criteria for rapidly deepening ACs

For this study, rapidly deepening cyclones are defined by normalizing the 24 h decrease in the central minimum pressure of the cyclone by the 95th percentile of observed cyclone deepening rates as follows,

$$\text{RD} = \frac{\Delta P_{\text{obs}}}{\Delta P_{\text{p95}}}, \quad (2.6)$$

where ΔP_{obs} denotes the change in central minimum MSLP over a 24-h period, and ΔP_{95} denotes the 95th percentile of observed cyclone deepening rates per 2.5° latitude bin. Surface cyclone tracks are obtained from the Crawford and Serreze (2016) identification and tracking algorithm applied to ERA5, with tropical cyclones filtered out by removing tracks that match tropical cyclone tracks from the IBTrACS database (Knapp et al. 2010) for either at least a consecutive 24 h period, or 60% of the cyclone's lifetime. Additionally, ACs in this study are filtered to cyclones that spend at least 24 h in the Arctic, with overall AC lifetime of at least 48 h.

There is a dynamical basis in normalizing by observed cyclone deepening rates. Figure 2.4 compares the Bergeron criteria and annual RD criteria per latitude, with a separate comparison against climatological 1000–500 hPa thickness and 850-hPa EGR. Low-mid tropospheric baroclinic instability is an important contributor to RD, and while the peak latitudes of the annual mean zonally averaged 850-hPa EGR and 1000–500 hPa meridional gradient of thickness gradient differ, both decrease with latitude poleward of 60°N . This generally matches the 95th percentile of analyzed cyclone deepening rates, which peaks near 47.5°N and decreases poleward.

Comparing the cyclone track densities of the portions of a cyclone's life cycle that exceed at least 1 RD units (RDU) (Fig. 2.5) with the equivalent calculation for Bergeron criteria (cf. Fig. 2.3), the midlatitude peak in RDU criteria is similar spatially but with a slightly lower peak than Bergeron criteria, as the RDU criteria exceeds Bergeron criteria between approximately 33°N and 52°N . The greatest differences between both metrics emerges in the Arctic with much greater cyclone counts that exceed RDU criteria. This finding motivates using RDU criteria instead of Bergeron criteria to examine RD ACs.

An advantage of using a normalization method to quantify RD is that there is an objective threshold defining when an RD episode starts and ends. This contrasts with compiling the most rapidly deepening cyclones in the entire Arctic basin, where defining when an RD

episode starts and ends is more subjective, despite being a quantitatively better metric for strictly identifying the most rapidly deepening ACs.

Cases are selected by filtering all ACs to those that exhibited lifetimes of at least 48 h, spent at least 24 h poleward of 66.6°N , and had at least one RD episode with the peak deepening rate poleward of 66.6°N . There is an 87% match between all ACs that exceed 1 RDU with the equivalent number of the top 24-h AC deepening rates. While not a perfect match, this is considered sufficient for the purposes of analyzing the beginning, peak, and end of an RD episode. Following this methodology, a total of 849 RD cases are obtained. For experiments that require partitioning by seasons, observed cyclone deepening rates used for normalization are changed to use data from that season, resulting in 263 RD cases for the summer season (June–August) and 178 RD cases for the winter season (December–February).

2.2.3 Storm-centered composites

To analyze relevant features to RD ACs such as TPVs in a cyclone-centered framework, cyclone-centered composites are created for all ACs that meet RD criteria using ERA5 data with 0.5° horizontal grid spacing. For TPV identification, composites of 2 PVU potential temperature and vorticity are created. To quantify mid-upper tropospheric forcing for ascent, composites of 500–700 hPa differential geostrophic absolute vorticity advection by the geostrophic are created, with a 500-km area-averaging radius applied. To quantify lower-tropospheric baroclinic instability, composites of 850-hPa EGR are created.

These composites are rotated using the average 250–850 hPa wind vector averaged within a 1000-km box surrounding the location of the minimum pressure of the surface cyclone, such that the average wind vector is always oriented to the right. This method is advantageous over rotating the composites using the cyclone motion vector as cyclones can exhibit rapidly fluctuating motion vectors especially during occlusion, which affects the location of nearby features influencing the AC. Additionally, using a tropospheric-averaged

motion vector has a dynamical basis, as the nearest TPV is expected to be located behind and slightly to the left of a cyclone with dominant upper-tropospheric forcing during its maximum growth rate, such that the maximum differential cyclonic vorticity advection would be centered near the cyclone of interest.

An important limitation is the sensitivity of results to the method of rotating the composites. The motion vector by which the composites are rotated was tested with various tropospheric layers and radii, with the 250–850 hPa wind averaged over a 1000 km box surrounding the AC producing the most robust and temporally consistent composite mean signatures of nearby features relevant to the AC, such as the composite mean upper-tropospheric trough magnitude and structure.

Where applicable, composite differences are tested for statistical significance at the 95% confidence level using a 10000 resample bootstrap test following the methodology of Rios-Berrios et al. (2016).

2.2.4 TPV matching

After data is converted to storm-centered coordinates and rotated as described in the previous section, the next step is to match ACs to nearby TPVs. While the easiest method to do so is by comparing the distance from the AC to the nearest TPV as identified by a minimum in 2 PVU potential temperature, this method does not account for the structure, size and asymmetry of the TPV, as well as where the location of the minimum 2 PVU potential temperature is within the vortex.

To account for this shortcoming, TPV shapes are identified in a 3000 km domain surrounding the AC using TPVTrack, which outputs TPV watershed basins. For each TPV basin, a mass centroid is computed based on the normalized 2 PVU potential temperature anomaly relative to ERA5 1991–2020 climatology and standard deviation. The closest mass centroid coordinate to the AC is identified as the closest TPV. Applying this approach results in nearly all identified TPV matches being situated behind and to the left of the

AC center relative to the deep layer motion vector. In contrast, matching TPVs using the 2 PVU potential temperature minima coordinates results in many more TPVs being identified well to the right of the motion vector, where they would not be expected to be dynamically relevant to the AC.

For the purposes of this study, there is no temporal tracking and matching of TPVs. TPVs are identified and matched at each individual time step analyzed, such that at any given time, the closest TPV to the AC is considered a match even if a tracking algorithm may not identify it as the same TPV throughout the course of an AC intensifying.

2.3 Model experiment setup

Experiments that perform numerical simulations use the Model for Prediction Across Scales (MPAS; Skamarock et al. 2012) version 8.0.1. MPAS is a fully compressible nonhydrostatic model using an unstructured Voronoi mesh. This unstructured mesh provides an advantage for global simulations over standard latitude-longitude grids in high latitudes by avoiding convergence into a singularity point at the North Pole. Another advantage of MPAS over regional models such as the WRF (Skamarock et al. 2005) is its capability for variable resolution meshes with a gradual refinement region, thus mitigating rigid boundary issues that exist with nested WRF domains (e.g., Skamarock et al. 2012; Park et al. 2014).

All MPAS simulations in this dissertation leverage its limited-area simulation capability (Skamarock et al. 2018) by extracting a regional mesh over the domain of interest for each experiment. Regional simulations offer benefits by reducing the computational expense of simulations and constraining the simulation within the domain of interest to the lateral boundary conditions (LBCs) input at various intervals. If the ICs and LBCs selected are from a reanalysis dataset, this can constrain forecast error growth relative to a global simulation. Nonetheless, limitations exist as spurious artifacts can arise in the regional domain boundary due to mismatches between the LBCs and the solution within the regional domain (Skamarock et al. 2018).

Parameterization	Scheme
Convection	New Tiedtke
Microphysics	Thompson
Land surface	Noah
Boundary layer	MYNN
Surface layer	MYNN
Radiation, LW	RRTMG
Radiation, SW	RRTMG

Table 2.1: MPAS v8.0.1 model configuration used for all simulations.

ICs and LBCs are obtained from the ERA5 reanalysis (Hersbach et al. 2020) at 0.25° horizontal grid spacing, except for the PL experiment, which is described in more detail in section 2.3.3, with LBCs updated at 3-hourly intervals. The full model configuration specifications are provided in Table 2.1.

2.3.1 TPV intensity modification algorithm

Experiments making use of MPAS include several simulations modifying the TPV intensity to quantify the impact of TPV intensity on the features of interest. A slightly modified version of the TPV intensity modification algorithm detailed in section 2.4 of Szapiro (2019) is used, where an upper-tropospheric heating or cooling tendency is applied to the TPVs at the beginning of each simulation. Recall from section 1.3 that upper-tropospheric potential temperature tendency is related to TPV intensity changes via the EPV tendency equation (Pedlosky 1998), as changes in EPV relate to changes in TPV intensity (Cavallo and Hakim 2009), TPV intensification is modulated by longwave radiative cooling (Cavallo and Hakim 2010, 2013), and the cold anomaly beneath the vortex core peaks in magnitude in the upper troposphere (Cavallo and Hakim 2010). Given these considerations, applying

an upper-tropospheric cooling tendency intensifies the TPV while an upper-tropospheric heating tendency weakens the TPV.

For each experiment, a control simulation was run alongside several simulations applying various temperature tendencies that are specified in each section below. These tendencies are applied for cells defined as being within the TPV, which is determined through a variable “inVortex” which is set to 0.0 for cells outside of the vortex and 1.0 for cells within the vortex. To determine where inVortex should be set to a non-zero value, and to prevent the vortex from completely dissipating in simulations designed to weaken the TPV, all simulations apply a linear interpolation to inVortex based on potential temperature on the 2 PVU surface. 2 PVU potential temperatures at or above a maximum threshold value selected for each case to encompass the outer edge of the TPV denotes an inVortex value of 0.0, and potential temperatures at or below a minimum value selected for each case to encompass the core of the TPV denote an inVortex value of 1.0. The vertical profile of the potential temperature tendency is then constructed for each cell as follows:

$$\text{tend}(z) = \begin{cases} 0 & \theta_z > \theta_{\text{trop}} \\ \text{tend} * \text{inVortex} & \theta_{\text{trop}} > \theta_z > \theta_{\text{trop}} - 10 \text{ K} \\ 0 & \theta_z < \theta_{\text{trop}} - 10 \text{ K}, \end{cases} \quad (2.7)$$

where $\text{tend}(z)$ is the vertical profile of the potential temperature tendency (tend), θ_z is the potential temperature value at vertical level z , and θ_{trop} is the 2 PVU potential temperature value at that cell. This algorithm is only applied for a period of time at the beginning of each simulation in order to spin up the TPVs, with the duration over which it is applied slightly different for each experiment, after which the control configuration is used for all simulations.

This methodology to modify TPVs results in random noise perturbations originating in areas of parameterized convection nearly immediately from the onset of the simulation, including locations well upstream of the domain of interest. These perturbations grow

nonlinearly with time, eventually propagating into the domain of interest such that differences between simulations can no longer be predominantly attributed to modifications to the TPVs, often occurring approximately 5 d into the simulations. This is known as “chaos seeding” (Ancell et al. 2018); chaos seeding is unavoidable within the Weather Research and Forecasting (WRF; Skamarock et al. 2005) model, and this apparently is the case for MPAS as well. The regional MPAS domain for each experiment was selected to greatly minimize the areas of chaos seeding observed when testing with a global domain, while ERA5 LBCs ensure consistent upstream synoptic features originating outside of the domain across all simulations.

The model configuration and methodology described in this section are used for the three modeling experiments in this dissertation, each described below in more detail.

2.3.2 August 2012 AC experiment setup

The control simulation for this case is initialized at 0000 UTC 4 August 2012. This allows for a 24 h window to modify the primary TPV associated with AC12, before AC12 exits Siberia and begins its peak RD episode. Figure 2.6 shows the MPAS regional domain used for this experiment, with mesh spacing of approximately 15 km.

Several sets of experiments are performed to quantify the impacts of the primary TPV (TPV1) and a separate TPV associated with the antecedent AC to the north of AC12 (TPV2) on the RD episode of AC12. The first experiment modifies TPV1 by applying potential temperature tendencies of -15 K day^{-1} (TPV1+) and -30 K day^{-1} (TPV1++) to intensify TPV1, and a temperature tendency of 10 K day^{-1} (TPV1-) to weaken the TPV1. In addition, another experiment is designed to weaken not just TPV1 but the broad upper-level trough that encompasses it, by applying a 20 K day^{-1} tendency to a larger domain surrounding the TPV (Trough-).

The second experiment is designed to test the impact of TPV2 on the evolution of AC12. As TPV2 becomes more easily distinguishable from the nearby 2 PVU potential

temperature field at 0300 UTC 4 August 2012, TPV2 is only modified over a 21 h window, between 0300 UTC 4 August and 0000 UTC 5 August 2012. This experiment applies potential temperature tendencies of -15 K day^{-1} (TPV2+) and -30 K day^{-1} (TPV2++) to intensify TPV2. Finally, to quantify the combined impact of both TPVs on the evolution of AC12, two simulations are performed with potential temperature tendencies of -15 K day^{-1} (Both+) and -30 K day^{-1} (Both++) applied to both TPVs.

Finally, the sensitivity of the TPV modification algorithm to the 2 PVU potential temperature thresholds selected is tested through several experiments varying the minimum and maximum 2 PVU potential temperature thresholds for TPV1. The control configuration uses a minimum value of 295 K and maximum value of 310 K. A “small” experiment decreases the maximum value to 305 K, a “large” experiment increases the maximum value to 315 K, and a “largest” experiment increases the minimum value to 300 K while retaining the maximum value of 315 K.

2.3.3 February 1987 PL experiment setup

Two control simulations are performed for this case, initialized at 0600 UTC and 1200 UTC 25 February 1987. This allows for at least 6–12 h lead time prior to the onset of more rapid cyclogenesis, and 24–30 h lead time prior to the cyclone becoming occluded and the end of the baroclinic deepening phase. Figure 2.7 shows the MPAS regional domain used for this experiment, with mesh spacing of approximately 15 km.

In addition to the objective of the control experiments to assess the deterministic forecast skill of the PL, another objective of this experiment is to assess the impact of IC uncertainty on the PL evolution and how varying TPV intensity affects these simulations. A 10-member IC uncertainty ensemble is available as part of ERA5, provided at a coarser 0.5° grid spacing and at 3-h temporal resolution. For both initialization times, a 10-member ensemble is run with these members used for ICs, while deterministic 0.25° ERA5 conditions still used for lateral boundary conditions.

The final objective of this experiment is to analyze the impact of TPV intensity on the PL evolution and whether the sensitivity of the PL evolution to TPV intensity exceeds that of sensitivity to IC uncertainty. For each forecast initialization, another 10-member ensemble is run with variable TPV intensity. For the 0600 UTC 25 February 1987 initialized ensemble, a -20 K day^{-1} potential temperature tendency is applied in the first 12 h of the simulation to intensify the TPV. For the 1200 UTC 25 February 1987 initialized ensemble, a 20 K day^{-1} potential temperature tendency is applied in the first 12 h of the simulation to weaken the TPV. The selection of a 12 h window to modify the TPV accounts for the short window between when the TPV exits Svalbard and when the PL begins to rapidly deepen. The 2 PVU potential temperature bounds used for inVortex are a minimum of 265 K and maximum of 285 K.

To quantify the sensitivity of the PL evolution to upstream features such as the TPV, ESA is applied for each of the ensembles described above. The response function J is the leading mode of empirical orthogonal function (EOF) of the MSLP field surrounding the ensemble mean cyclone location, which describes the leading mode of variability of MSLP, selected at 1200 UTC 26 February 1987 when the PL reaches its peak intensity.

As discussed in section 1.5.2, some PLs are very sensitive to the horizontal resolution of the simulation (McInnes et al. 2011). Simulations with 3.75-km mesh spacing showed a nearly identical PL track and intensity to those with 15 km mesh spacing. In addition, the 2 PVU potential temperature field in the 3.75-km simulation exhibited widespread small-scale noise affecting the quality of the TPV intensity modification algorithm. Accounting for these reasons, and in part for computational resource efficiency, all simulations for this experiment use 15 km mesh spacing.

2.3.4 February 2021 CAO experiment setup

The control simulation for this experiment is initialized at 0000 UTC 8 February 2021. This lead time was selected to allow sufficient time to modify both TPVs at the onset of

the CAO and prior to the TPVs' interaction, while skillfully resolving the TPV and CAO evolution at several days lead time. Figure 2.8 shows the MPAS regional domain used for this experiment, with mesh spacing of approximately 15 km.

Seven simulations modifying the TPV intensity are performed to quantify the impact of TPV intensity on the CAO evolution. Alongside a control simulation, three simulations weaken the TPV by applying potential temperature tendencies of 5 K day^{-1} , 10 K day^{-1} and 15 K day^{-1} , and three simulations intensify the TPV by applying potential temperature tendencies of -5 K day^{-1} , -10 K day^{-1} and -15 K day^{-1} . The 2 PVU potential temperature bounds used for inVortex are a minimum of 275 K and maximum of 300 K.

The TPV intensity modification algorithm is applied over a 24-h period between 1200 UTC 8 February 2021 and 1200 UTC 9 February 2021 in each simulation in order to spin up the TPVs, after which the control configuration is used for all simulations. This computation is performed over cells in a domain centered over south-central Canada, with temporal modifications to the domain applied to avoid modifying unrelated PV filaments. In particular, a lobe of one of the TPVs of interest splits into a separate vortex. In simulations that modify this lobe during and following the TPV split, the vortex that forms because of the split is a source of large forecast sensitivity on the synoptic scale.

The domain within which TPVs are modified for this experiment (Fig. 2.9) was subjectively selected in an attempt to avoid modifying the previously noted vortex lobe. For comparison purposes, simulations that include the lobe in the calculation will also be discussed. This domain was further selected to encompass relevant synoptic features, primarily the TPVs over southern Canada, the poleward ridge that impacts their zonal progression, and the subsequent cold air advection and cyclones over the contiguous U.S.

As with the other NWP modeling experiments, the source and quality of ICs is an important source of uncertainty. For this case study, a comparison of simulations initialized with Global Forecast System (GFS) NWP model and ERA5 ICs showed the latter had the

closest synoptic evolution of the CAO to what was observed, supporting the use of ERA5 ICs.

To quantify the sensitivity of the CAO to the TPV intensity and evolution, two sets of ESA calculations are performed. For both sets, the response function J is defined as 2-meter temperatures averaged over the southern Great Plains domain (Fig. 2.10), intended to quantify both the southward extent and magnitude of the CAO. The valid time of J is 1200 UTC 15 February 2021, at the coldest time of the CAO in the southern Great Plains. First, ESA is calculated for the variable TPV intensity ensemble described above, as it would be expected that the TPV intensity at 24 h forecast lead time would show as a region of forecast sensitivity. Second, to compare the results of this experiment to operational forecast uncertainty, ESA is calculated for the 51-member operational ECMWF ensemble at various lead times preceding the peak of the CAO.

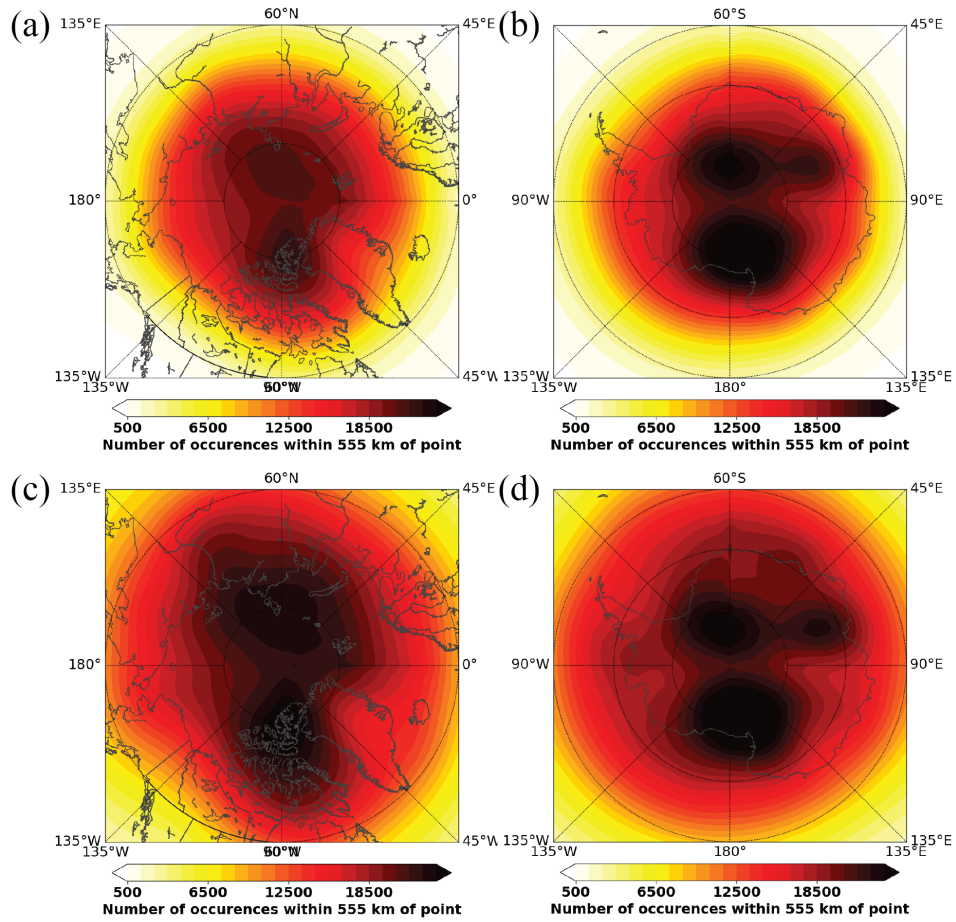


Figure 2.1: Number of TPV occurrences within 555 km of a point using (a, b) criteria from Cavallo and Hakim (2009), and (c, d) criteria from Szapiro and Cavallo (2018). Plotted are the (a, c) Northern Hemisphere, and (b, d) Southern Hemisphere.

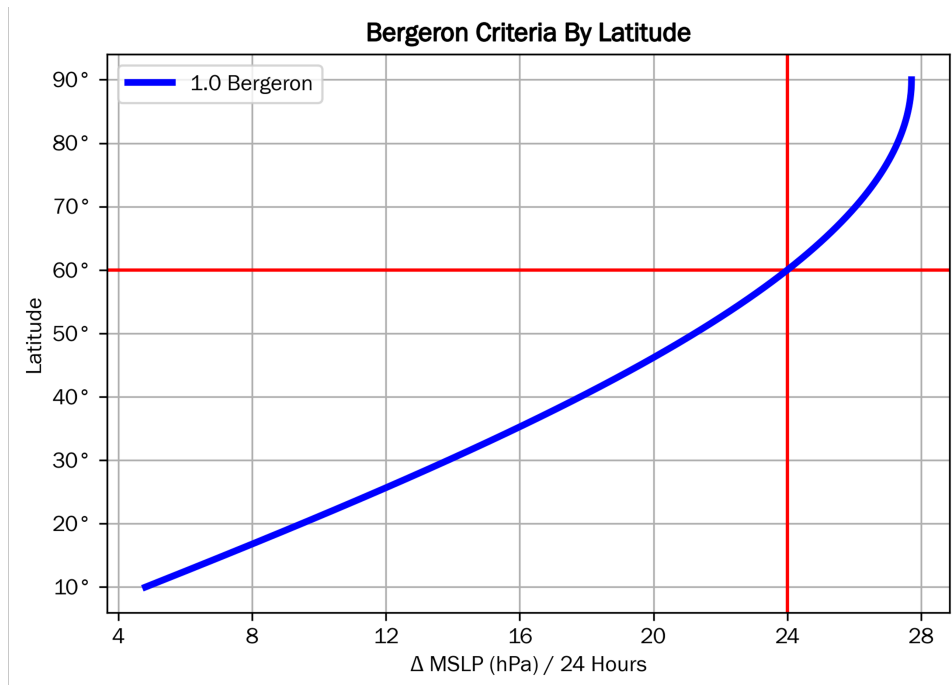


Figure 2.2: Bergeron criteria as a function of latitude in the North Hemisphere. Blue line denotes a value of 1.0 Bergeron. Red lines denote the default 60°N reference latitude and associated 24 hPa per 24 h threshold.

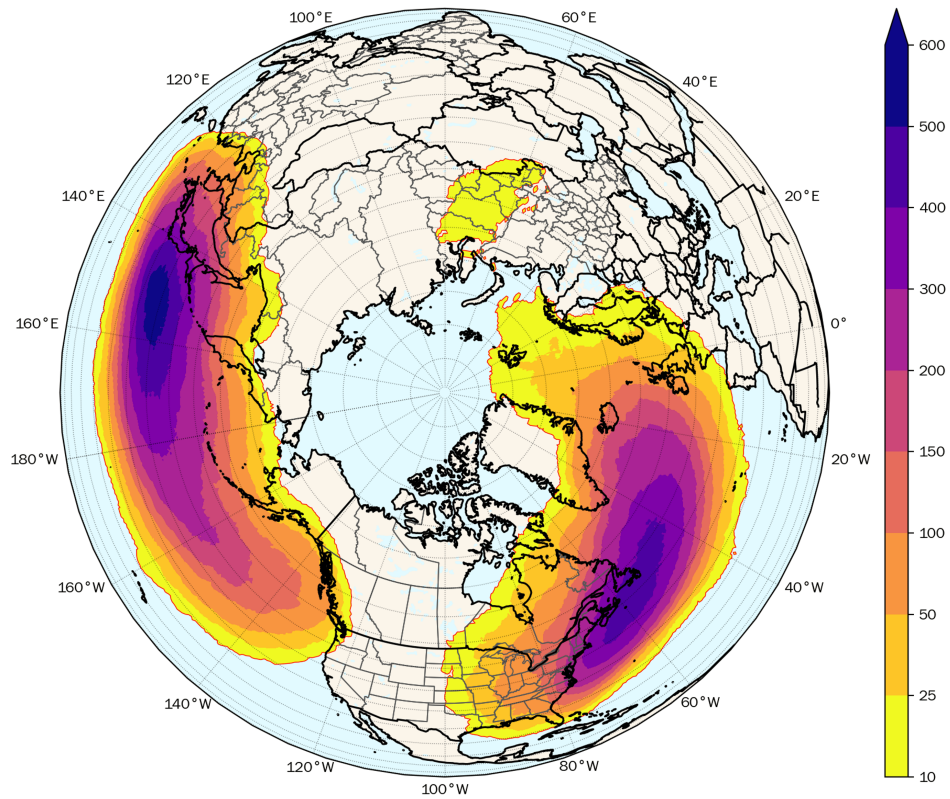


Figure 2.3: 500-km track density of the portion of cyclone tracks where the 24 h deepening rate exceeds 1 Bergeron, using 1984–2016 ERAI cyclone tracks.

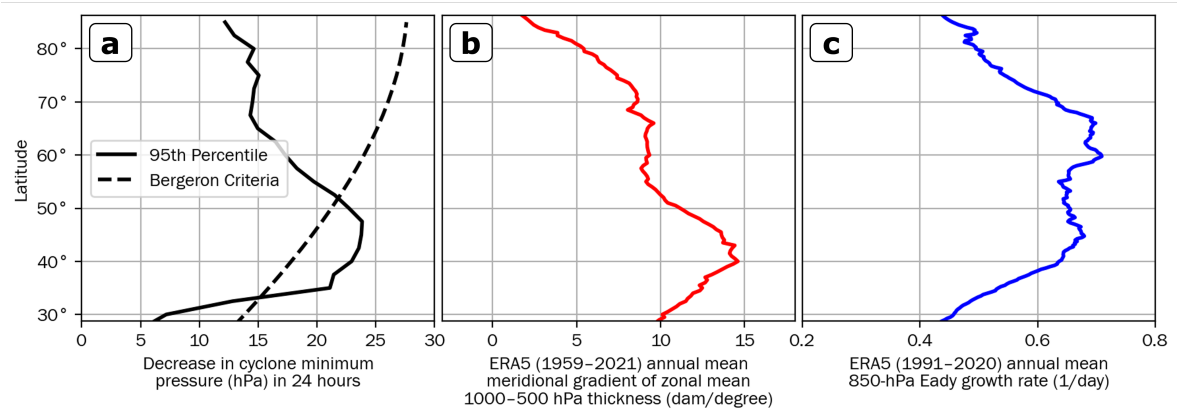


Figure 2.4: North Hemisphere comparisons of zonally averaged (a) Bergeron criteria (dashed line) and 95th percentile of observed cyclone deepening rates per 2.5° latitude bin (solid line), (b) annual mean meridional gradient of zonal mean 1000–500 hPa thickness (dam deg^{-1}), and (c) annual mean 850-hPa EGR (day^{-1}).

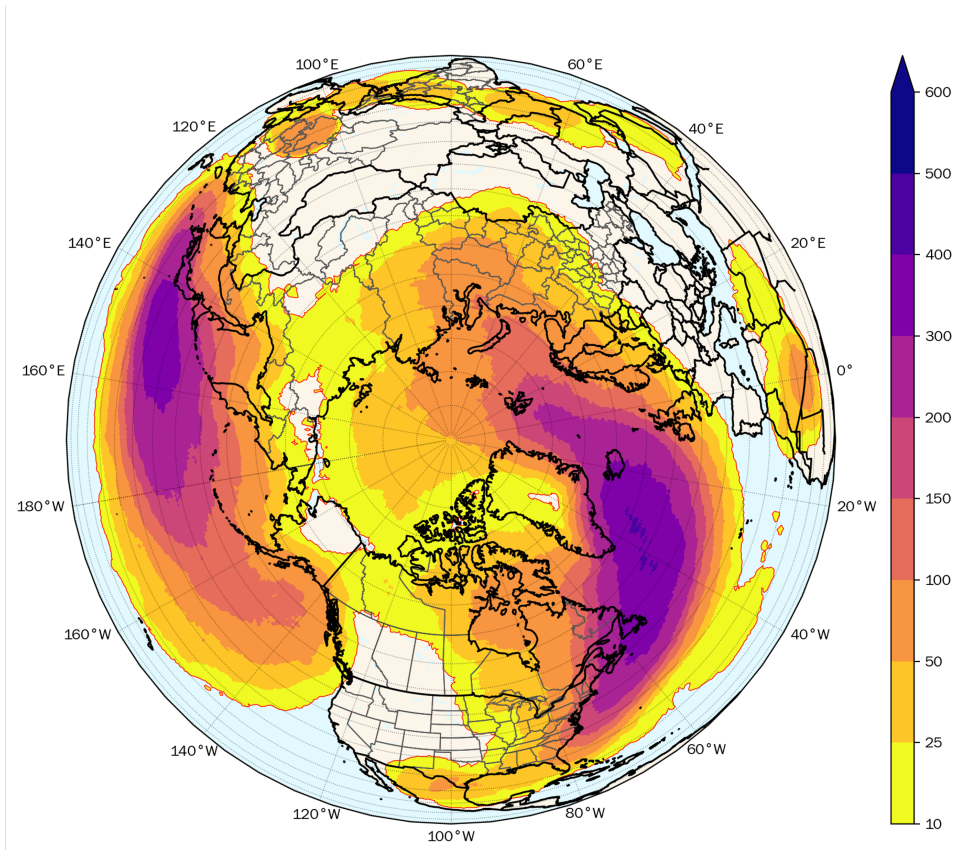


Figure 2.5: Same as Fig. 2.3, but for the 95th percentile threshold of observed cyclone deepening rates.

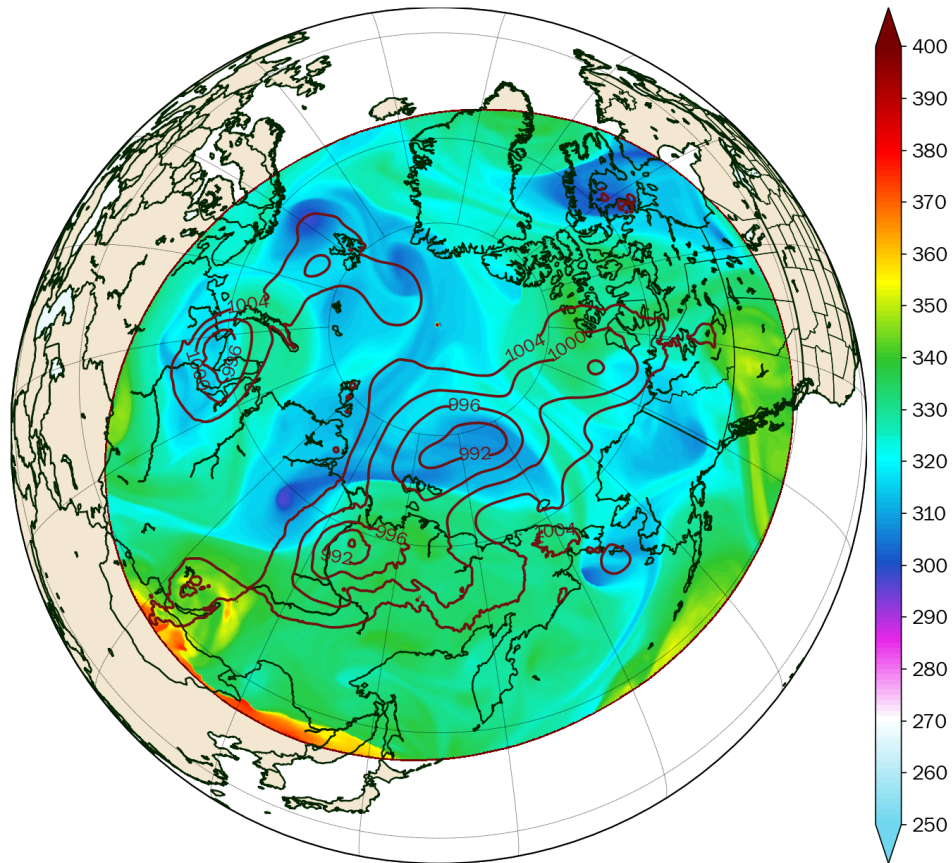


Figure 2.6: MPAS regional domain used for the 2012 “Great Arctic Cyclone” simulations. Within the domain, ERA5 ICs are shown for 2 PVU potential temperature (K, fill) and MSLP (every 4 hPa, brown contours). Mesh spacing within this domain is approximately 15 km.

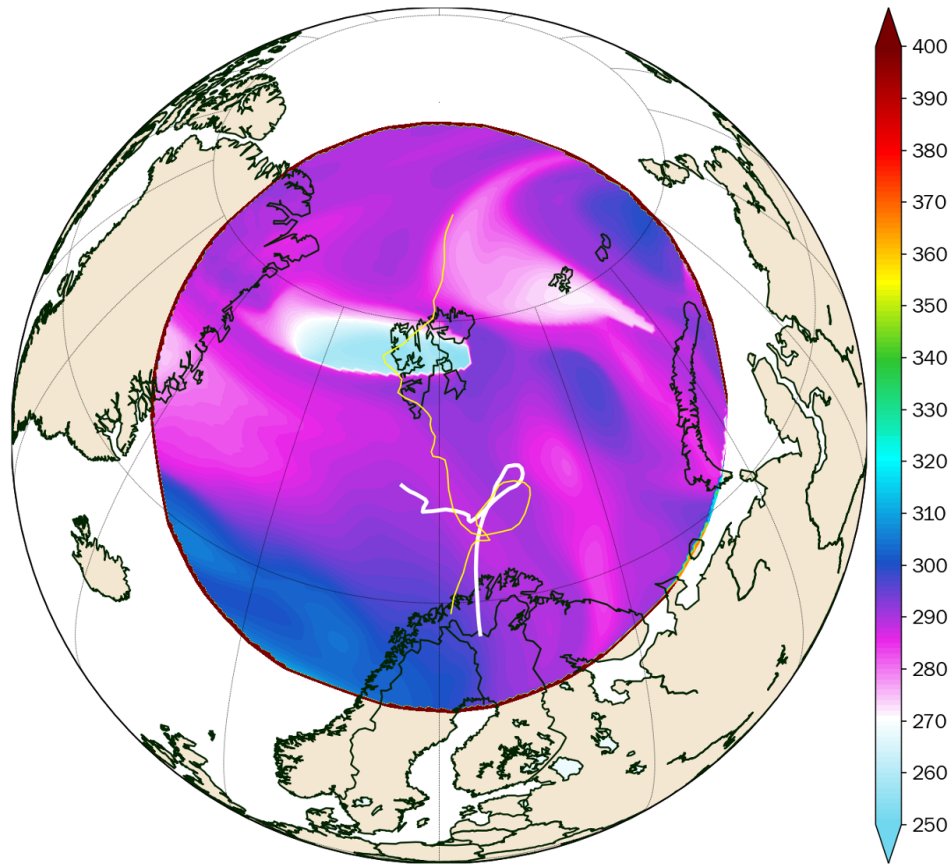


Figure 2.7: MPAS regional domain used for the February 1987 PL simulations. Within the domain, ERA5 ICs are shown for 2 PVU potential temperature (K, fill). The solid white line denotes the ERA5 PL track based on hourly MSLP minima, and the thin yellow line denotes the ERA5 TPV track based on hourly 2 PVU potential temperature minima. Mesh spacing within this domain is approximately 15 km.

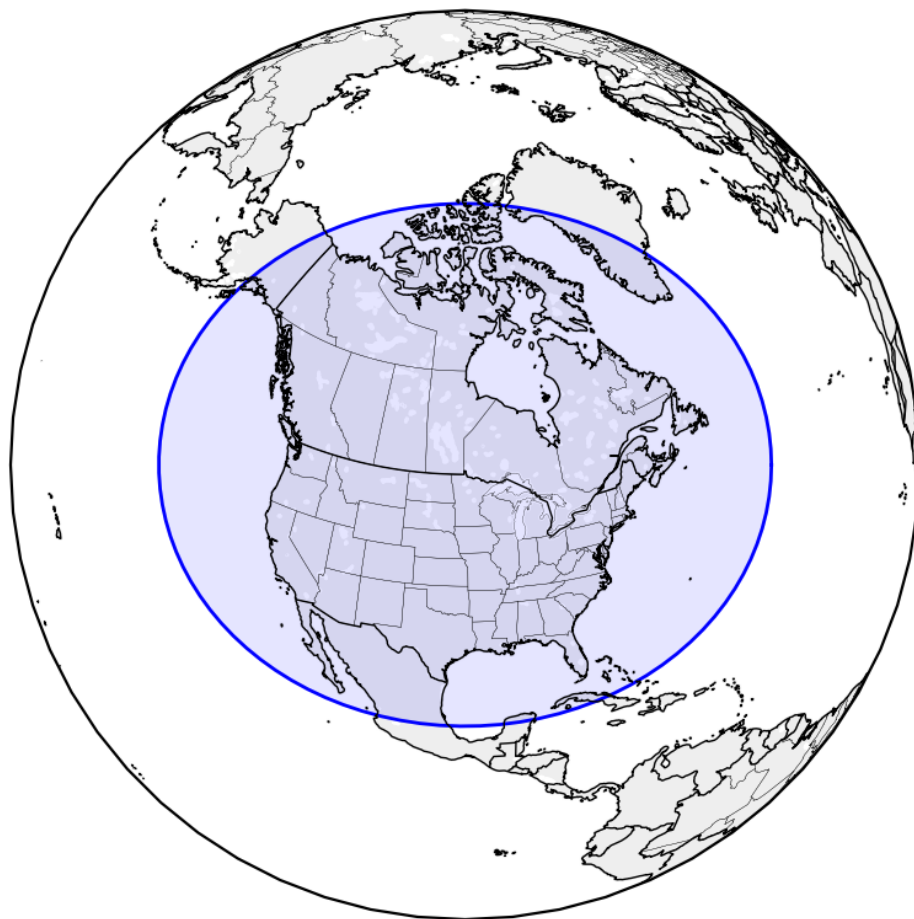


Figure 2.8: MPAS regional domain used for the February 2021 CAO simulations. Mesh spacing within this domain is approximately 15 km.

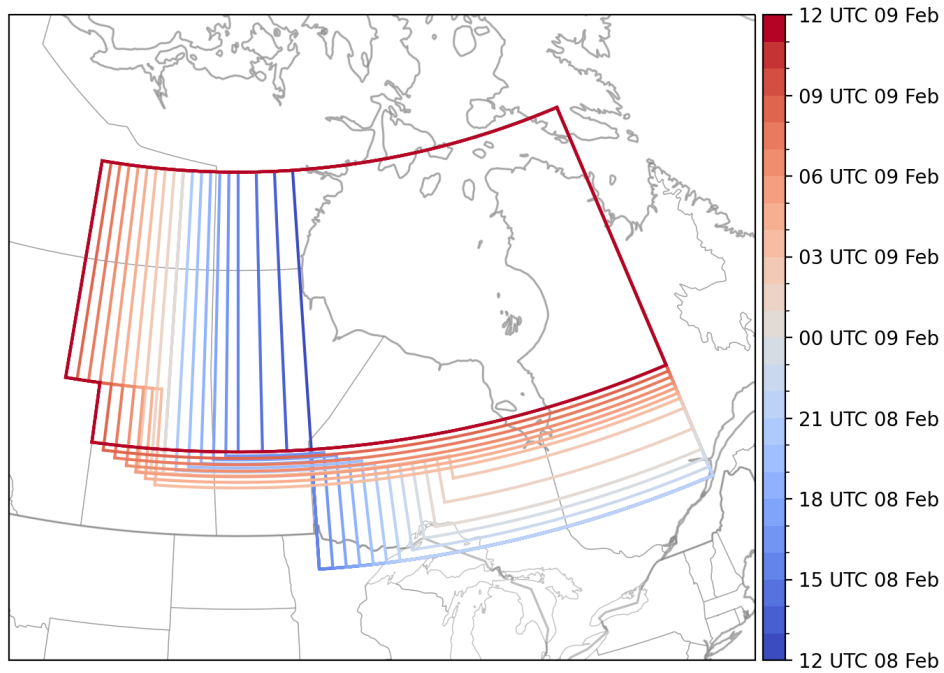


Figure 2.9: Domain within which the TPV intensity modification algorithm is applied to as a function of time. The temporal adjustments in this domain encompass both TPVs of interest, while avoiding a separate PV lobe to the southwest and attempting to separate one of the TPVs of interest from a fragment that splits into a separate TPV.

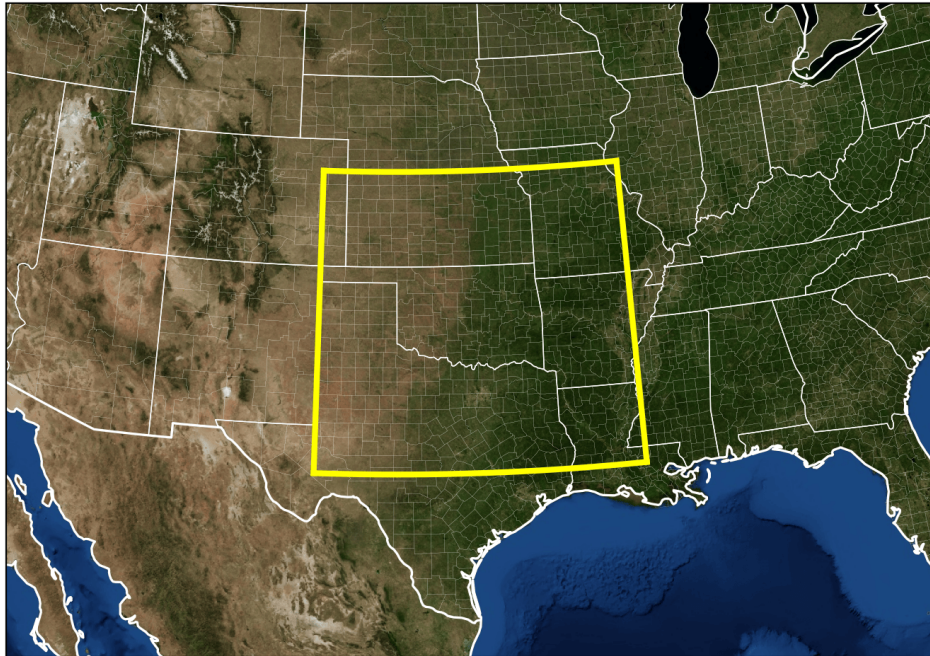


Figure 2.10: Southern Great Plains domain used for several analyses in Chapter 5.

Chapter 3

Characteristics of rapidly deepening Arctic cyclones

This chapter aims to improve our understanding of systematic characteristics of rapidly deepening ACs, focusing on their rapid deepening (RD) episodes, defined as the portion of the life cycle of an AC where it exceeds RD criteria discussed in section 2.2.2. Their associations with nearby TPVs and their baroclinic instability as approximated by 850-hPa EGR is analyzed, with RD cases partitioned by season, region, and time of cyclogenesis relative to the onset of RD. All spatial composites of RD cases, centered on the AC, are computed following the methodology described in section 2.2.3, rotated in the direction of the 1000 km averaged 250–850 hPa averaged wind vector.

There are several factors to consider to consider when interpreting these results:

- While the RD criteria for annual ACs is based on an annual latitudinal climatology of cyclone deepening rates, the latitudinal climatology is re-calculated for all analyses involving summer and winter cases using a seasonal latitudinal climatology of cyclone deepening rates.
- Out of all ACs that meet annual RD criteria, 20 ACs exhibit multiple RD episodes. All analyses in this chapter only include the first RD episode for these ACs to get maximum variety between cases within each composite.
- Unless otherwise stated, all references to “anomalies” in this chapter refer to anomalies relative to ERA5 climatology from 1991–2020, constructed using a 60-day centered window with triangular weights.

3.1 Rapidly deepening AC climatology

There are a total of 849 cases that meet the RD annual criteria. Their spatial distribution is represented by occurrence frequency (Fig. 3.1a), calculated by the percent of 6-h time steps between 1979–2022 a grid point is located within 500 km of the portion of the track of an AC where RD value surpasses 1.0. Occurrence frequency is largest in the North Atlantic Ocean, peaking between Iceland and Svalbard, and gradually decreases towards the Arctic coast along eastern Siberia and North America. This region of maximum occurrence frequency aligns with a climatological annual maximum in 850-hPa EGR and the meridional gradient of 1000–500 hPa zonal mean thickness (not shown).

Annual cyclogenesis occurrence frequency (Fig. 3.1b) of ACs that meet RD criteria is largest along the eastern coast of Greenland, with a maximum off the coast of east-central Greenland. This region is frequently in the marginal sea ice zone (MIZ), likely providing a source of low-level baroclinicity that can aid in cyclogenesis. A second, smaller maximum in cyclogenesis occurrence frequency extends along the Arctic coast over eastern Siberia and northern Alaska, likely associated in part with the AFZ.

The temporal evolution of the composite AC structure during the RD episode can be assessed by comparing the “start” of the RD episode when the RD value first reaches 1.0, the “peak” of the RD episode as defined by the center time of the 24-h window with the greatest MSLP decrease, and the “end” of the RD episode when the RD value falls below 1.0. Composite 2 PVU potential temperature and MSLP at the beginning of the RD episode (Fig. 3.2a) shows a large-scale trough upshear of the AC, with a 2 PVU potential temperature minimum indicative of the composite mean TPV signature situated upstream and to the left of the AC, such that the developing AC is situated downstream of the TPV. A broad ridge extends to the right of the AC, with the maturing AC embedded in a broad 1000–500 hPa thickness gradient (not shown). There is a broad signature of an antecedent AC downstream and left of the developing AC, although large spatial variability of the antecedent AC within the composite likely precludes a stronger signal in the composite mean.

As the RD episode reaches its peak (Fig. 3.2b), the composite mean TPV signature becomes more pronounced with a larger negative 2 PVU potential temperature anomaly shifting closer to the AC, with any antecedent AC merging into the newly dominant AC. This does not necessarily imply that the TPV intensity is stronger than at the start of the RD episode, however, as an increasing magnitude of the composite 2 PVU potential temperature anomaly can also signify less spatial variability in the TPV location within the case distribution. As the RD episode ends (Fig. 3.2c), a cyclonic PV hook signature becomes strongly evident in the composite mean, as the maturing surface cyclone becomes vertically aligned with the TPV and the maximum baroclinicity is displaced downstream and to the right of the AC.

The composite TPV structure during the RD episode can be alternatively assessed by compositing the shape of the TPV basin closest to the AC at each step of its RD episode. The closest TPV is identified through iterating over the mass centroid of each TPV basin, calculated with 2 PVU potential temperature normalized anomaly, within a 3000 km domain surrounding the AC. As this is a diagnostic analysis, the closest TPV to the AC during the start of the RD episode may not necessarily be the same TPV at the end of the RD episode. At the beginning of the RD episode (Fig. 3.2d), the composite closest TPV basin is situated upstream and to the left of the AC, but with pronounced spatial variability as peak gridpoint frequency within a TPV basin is only 35%. As the RD episode progresses (Fig. 3.2e-f), spatial variability in the location of the closest TPV decreases as its composite location shifts closer to the surface cyclone, eventually becoming vertically aligned with the AC.

The temporal evolution of the composite mean 500-hPa geopotential height (Fig. 3.3a) and 2 PVU potential temperature (Fig. 3.3b) fields at the peak of the RD episode can be quantified by subtracting the start from the peak of the RD episode. As the AC reaches its peak deepening rate, 500-hPa geopotential heights downstream of the composite mean trough axis decrease over and downstream of the AC. Dynamically, geostrophic cyclonic

vorticity advection is associated with forcing for negative height tendency per the QG height tendency equation. The spatial extent of 2 PVU potential temperature decrease is smaller than the 500-hPa geopotential height decrease, with the maximum decrease situated just upstream of the AC. This reflects the closest TPV mutually interacting with the deepening surface cyclone and wrapping around the surface cyclone.

3.2 Summer and winter composites

Seasonal differences of RD cases are examined by partitioning all cases into two groups. The first group consists of “summer” cases, defined as cases where the peak deepening rate of an AC, or the peak deepening rate in the first RD episode of an AC that has multiple RD episodes, occurs during the months of June, July, or August. The same criteria is used to define “winter” cases during the months of December, January, or February. After recomputing RD criteria for each season based on its latitudinal climatological cyclone deepening rates, there are a total of 263 cases during the summer, and 178 cases during the winter.

The spatial distributions of RD case tracks and cyclogenesis locations exhibit substantial seasonal variability. Whereas summer cases (Fig. 3.4b) are predominantly spread across the Arctic Ocean into northern Eurasia, winter cases (Fig. 3.4c) are more heavily concentrated in the Greenland Sea and Fram Strait, with lower occurrence frequency of RD cases extending into the Pacific side of the Arctic Ocean. This variance is in part explained by the poleward shift of the polar jet and decreased midlatitude baroclinicity in summer relative to winter. Cyclogenesis in the summer (Fig. 3.4e) is distributed mostly evenly across the Arctic with slight maxima along the Arctic coastlines, likely associated in part with enhancement along the AFZ. Cyclogenesis in the winter (Fig. 3.4f) is more heavily concentrated off the eastern coast of Greenland, along the climatological sea ice edge.

3.2.1 TPV proximity comparison

The TPV signature in the summer 2 PVU potential temperature composite throughout the RD episode (Fig. 3.5d,e,f) is more compact and situated closer to the AC center than in winter cases (Fig. 3.5g,h,i). Comparing composite means of 2 PVU potential temperature alone, however, does not necessarily indicate whether the TPVs in one group are significantly closer to the AC center than in another group. This comparison is better quantified by comparing distributions of the AC distance from the closest TPV mass centroid between composites. Fig. 3.6 compares the distance from the AC to the nearest TPV between the start, peak, and end of the RD episode between annual, winter, and summer cases. Summer cases exhibit a significantly closer TPV to the AC throughout the RD episode than winter cases, although this difference decreases with time such that by the end of the RD episode, it is significant at the 95% level but no longer significant at the 99% level. The TPV-AC distance distribution at the end of the RD episode is much more positively skewed for winter than summer cases, which suggests proximity to the closest TPV may be less of an important factor for rapid deepening in winter relative to summer cases.

Seasonal differences in AC proximity to the closest TPV are also a function of AC location. For the annual, summer, and winter case lists each, the distribution of AC distance to the closest TPV at the time of the peak deepening rate is partitioned into a “close” category, defined as an AC-TPV distance below the bottom quartile of this distance distribution, and a “far” category, defined as a AC-TPV distance above the upper quartile of this distance distribution. Figure 3.7 shows the difference in occurrence frequency for cyclone location density of ACs at the peak of their RD episode for the close TPV group minus the far TPV group. For both summer and winter seasons, cases in the central Arctic Ocean on the Russian side of the North Pole, and northern Nunavut, are more likely to be associated with TPVs in the close category than the far category, while cases from the Greenland Sea into the Eurasian coast are more likely to be associated with TPVs in the far category than the close category. As RD cases in the North Atlantic region are more common in the winter

than summer, this is linked with the tendency for summer cases to be associated with a closer TPV than in the winter. These summer results follow the same pattern as results from Gray et al. (2021) (their Fig. 7), where intense ACs are more likely to be matched with TPVs away from the Arctic coast than near the coastline.

3.2.2 EGR comparison

Repeating the prior analysis of partitioning AC-TPV distances into lower and upper quartile groups is repeated for both 850-hPa 500 km area-averaged EGR and EGR normalized anomaly. Differences in AC location occurrence frequency at the peak of the RD episode for the high minus low EGR groups (Fig. 3.8a-c) show ACs that rapidly deepen over the central Arctic Ocean tend to be associated with lower EGR than ACs that rapidly deepen over the Arctic coastlines and the North Atlantic. This result is to be expected, however, as climatological EGR is lower over the cold and dry central Arctic relative to the gradient near the Arctic coast and the North Atlantic between the cooler Arctic airmass and the warmer and more moist lower latitudes.

Repeating this analysis for EGR normalized anomaly (Fig. 3.8d-f), however, reveals a somewhat different pattern. Summer cases over the central Arctic Ocean are more frequently associated with higher EGR, and EGR normalized anomaly, than summer cases along the Arctic coastlines and the North Atlantic. Winter cases, however, are more frequently associated with higher EGR normalized anomaly over the Barents and Kara seas, and a lower EGR normalized anomaly over the Fram Strait and Greenland Sea.

The same type of analysis as shown with Fig. 3.6 is used for comparing EGR normalized anomaly between seasons (Fig. 3.9). Annually, rapidly deepening ACs exhibit no significant differences in EGR normalized anomaly between different phases of the RD episode. This lack of temporal trend throughout the RD episode for annual cases, however, obscures opposing temporal trends by season. Whereas EGR normalized anomaly decreases throughout the RD episode for summer cases, it increases throughout the RD

episode for winter cases. This result was not expected, and may be associated with overall greater deepening rates and lower minimum MSLP for winter cases than summer cases.

A more direct linkage between AC-TPV distance and EGR normalized anomaly is shown in Fig. 3.10. For summer cases (Fig. 3.10a), there is a statistically significant correlation such that cases farther from the closest TPV are associated with a more positive normalized EGR anomaly, and vice versa. For winter cases (Fig. 3.10b), however, there is no correlation between the AC-TPV distance and EGR normalized anomaly. This indicates that proximity to the closest TPV plays a greater role in the rate of intensification of summer cases than winter cases, as baroclinicity is generally lower during summer than winter.

There is a statistically significant difference in total AC lifetimes of RD cases between summer and winter. Summer ACs that meet RD criteria have an average lifetime of 6.4 days, with an interquartile range (IQR) of 3.8 to 8.3 days, while winter ACs that meet RD criteria have an average lifetime of 5.1 days, with an IQR of 3.5 to 6.0 days. This difference becomes most pronounced in the upper tail of the distribution, with the 95th percentile of AC lifetimes at 21.1 days for summer cases, compared to just 12.0 days for winter cases. This result for rapidly deepening ACs aligns with climatologies of intense ACs that found intense ACs on average have longer lifetimes in summer than winter (e.g., Vessey et al. 2020).

3.3 Regional composites

Spatial differences can be generally approximated by partitioning ACs into two basins, an “Atlantic” basin (Fig. 3.11b) and a “Pacific” basin (Fig. 3.11a). Cases are grouped into the Atlantic basin if their peak deepening rate occurs equatorward of 85°N, east of 40°W, and west of 80°E. This domain encompasses the region between the northeast Atlantic Ocean north of Iceland into Svalbard, and the Fram Strait, Kara Sea, and Barents Sea, and follows the climatological maximum in low-level baroclinicity. The remaining cases are

classified into the Pacific basin, encompassing the remainder of the Arctic Ocean and the northern Eurasian and North American coastlines. There are a total of 569 cases in the Atlantic basin, and 280 cases in the Pacific basin. Both basins share a common cyclogenesis region in the eastern coast of Greenland, though Pacific cases (Fig. 3.11c) are more heavily weighted towards northern Greenland, with a secondary maximum in cyclogenesis frequency over the Arctic coast, and Atlantic cases are more heavily weighted towards east-central Greenland.

During the winter, 73% of cases occur in the Atlantic basin, and 27% occur in the Pacific basin. In contrast, during the summer, 37% of cases occur in the Atlantic basin, and 63% occur in the Pacific basin. While EGR is significantly larger for Atlantic cases than Pacific cases, an expected result given EGR climatological differences between both basins, EGR normalized anomaly (Fig. 3.12) is significantly more positive for Pacific cases than Atlantic cases at the start and peak of the RD episode. A possible explanation for this discrepancy is that given the lower climatological EGR in the Pacific basin relative to the Atlantic basin, anomalously strong baroclinic instability is necessary to support rapid deepening. This may be associated with variability in the magnitude of the AFZ, or perhaps an association of AC genesis along the Siberian coastline with anomalous poleward warm and moist air advection. By the end of the RD episode, there is no significant difference in EGR normalized anomaly between Pacific and Atlantic cases.

There is a significant difference in the lifetimes of ACs despite the onset of their RD episode between the Atlantic and Pacific basins, with a much longer duration preceding RD for Pacific cases than Atlantic cases. This discrepancy is elaborated more on in section 3.4. However, unlike the seasonal difference composites, there is no significant difference in overall AC lifetimes between Atlantic and Pacific cases.

3.4 Onset of rapid deepening

Not all ACs begin their RD episode at similar times relative to the overall AC lifecycle. Cases can be partitioned into two groups based on the time of the onset of an AC's first RD episode relative to the time the cyclone is first identified. For the purpose of this analysis, this time will be referred to as the time of "cyclogenesis", though as discussed in section 2.1.3, the definition of cyclogenesis varies between different cyclone identification algorithms. Out of the 849 annual RD cases, 450 cases (53% of all cases) are first identified at the same time as the onset of the first RD episode, and are grouped into the "onset" category. A total of 232 cases (27% of all cases) are first identified at least 24 h or more before the onset of the first RD episode, and are grouped into the "24 hour" category.

The composite difference in MSLP normalized anomaly between both groups (Fig. 3.13a) shows a significantly more positive MSLP normalized anomaly over the AC for the onset composite than the 24-hour composite, which is an expected result as a well-defined cyclone does not exist for the onset group at the beginning of the RD episode. There is a broad signal of an antecedent AC in both the 24 hour and onset composites, although this signal is slightly, but not significantly, more pronounced in the onset composite. 2 PVU potential temperature normalized anomaly (Fig. 3.13b) is higher over the AC center in the onset composite, indicative of a more distant TPV to the cyclone at the beginning of the RD episode, than the 24-hour composite. While not shown here, at the beginning of the RD episode, the onset group has significantly (99% level) higher EGR normalized anomaly and a greater distance from the nearest TPV relative to the 24-hour group, though there are no significant differences in either comparison by the end of the RD episode. This may indicate that a nearby presence of a TPV plays a greater role in the onset of the RD episode for ongoing ACs than newly forming ACs that rely strongly on strong low-mid level baroclinicity to organize.

Spatial differences in cyclone locations at the onset of RD between the onset group and the 24-hour group are not surprisingly heavily weighted towards the onset group both

annually (Fig. 3.14a) and in the summer (Fig. 3.14b), given the much larger sample size in the onset group. However, annual cases show a slight preference towards the 24-hour group near the North Pole, while in the summer, two regions show a slight preference towards the 24-hour group, one along the northern coast of eastern Siberia, and another extending from Franz Josef Land into the Laptev Sea.

The regions that show a slight preference for the 24-hour group notably overlap with one of two TPV occurrence frequency maxima in the Northern Hemisphere, the other being over northern Nunavut. Cavallo and Burg (2024) found that while over 90% of ACs are associated with at least one TPV, about 48% of ACs are associated with two or more TPVs, which decreases further to 24% of ACs associated with three or more TPVs. One possibility is that the 24-hour group is more likely to be associated with multiple TPV interactions, although this hypothesis has not been tested as part of this study. Another possibility is that ACs in the onset group may be more likely to be associated with a decaying antecedent AC and asymmetric TPV, such that a lobe of the associated TPV forces a new surface cyclone to develop and quickly deepen, or with a separate TPV infringing upon the baroclinic zone associated with a decaying AC post-occlusion and forces renewed cyclogenesis.

3.5 Rapidly deepening vs. ordinary ACs

Ordinary cyclones are defined for this analysis as ACs that fail to meet RD criteria, but exhibit a peak minimum MSLP decrease of at least 4 hPa over a 24 h period. There are a total of 3956 ordinary AC cases. For computational efficiency, AC distances to the closest TPV for this analysis are calculated based on the coordinate of the minimum 2 PVU potential temperature, as opposed to the coordinate of the TPV mass centroid.

Notably, there is no significant difference in AC-TPV distance between rapidly deepening ACs and ordinary ACs (Fig. 3.15). The closest TPVs to rapidly deepening ACs do have a slightly larger TPV amplitude, defined as the difference in 2 PVU potential temperature between the TPV center and the outer edge of the TPV basin, than ordinary ACs, although

this difference is not statistically significant (not shown). However, rapidly deepening ACs do have significantly (99% level) lower 2 PVU potential temperature within the TPV center (not shown), and more positive EGR and EGR normalized anomaly, than ordinary ACs. A conclusion from this finding is that while TPVs do play an important role in rapid deepening, the presence of a close TPV to the AC is not a necessary condition for an AC to rapidly deepen.

3.6 Discussion

This chapter examined systematic characteristics of rapidly deepening ACs, comparing differences between seasons, regions in the Arctic, time of RD onset relative to time of cyclogenesis, and rapidly deepening and ordinary ACs. Overall results are in close agreement with studies that examined the spatial climatology and characteristics of intensifying ACs from a case list determined from the most intense ACs (e.g., Gray et al. 2021; Vessey et al. 2022; Croad et al. 2023), despite a different case selection and compositing methodology, but with some unique insights pertaining to the role of TPV proximity.

Returning to the hypotheses motivating this chapter, hypothesis A1 expected summertime cases to be associated with stronger forcing from a nearby TPV, longer lifetimes, and closer proximity to the nearest TPV during the RD episode relative to winter cases. Summer cases were found to have longer lifetimes and closer proximity to the nearest TPV, in addition to lower EGR and EGR normalized anomaly, than wintertime cases.

The QG omega equation is one metric that can quantify synoptic forcing for ascent via differential cyclonic vorticity advection. Comparing composites of QG omega term A for annual (Fig. 3.16a), summer (Fig. 3.16b), and winter (Fig. 3.16c) cases shows that while the composite mean maximum contribution for forcing for ascent from term A of the QG omega equation is maximized over the surface cyclone center, as would be expected, the composite mean magnitude is smaller for summer cases than winter cases. One potential

explanation is the substantial amount of spatial smoothing applied to the QG omega calculation, as this metric is often computed on coarse datasets. As TPVs are sub-synoptic scale phenomena, and spatial composites in Fig. 3.5 show the winter composite mean 2 PVU potential temperature has a broader negative anomaly than the summer composite, it is possible that the smoothing traditionally applied to QG forcing is sufficient for synoptic-scale analysis, but under-represents the peak magnitude of differential cyclonic vorticity advection for sub-synoptic scale TPVs. This may also affect the composite mean peak magnitude of term A where one composite has greater variability in TPV location relative to the AC than the other.

A noteworthy result is that winter cases in the Barents and Kara Seas are more likely to be associated with anomalously high EGR than RD cases in the Greenland Sea and Fram Strait. The role of the sea ice edge in facilitating rapid deepening in this region has not been examined as part of this study, and may be an opportunity for future work.

Separating rapidly deepening ACs into basin also yields information about differing characteristics by region. Atlantic cases are more frequent in the winter months than summer months, while Pacific cases are more frequent in the summer months than winter months. Atlantic cases are associated with larger EGR, but a less positive EGR normalized anomaly, than Pacific cases, and there is no significant difference in the AC-TPV distance between summer and winter cases throughout the RD episode. These results mostly do not align with the expectation from hypothesis A2, which may in part be explained by the stronger dependence on AC-TPV distance on variable location within the basin than broad subjectively defined basins.

Comparing ACs whose RD onset coincided with cyclogenesis to ACs that existed at least 24 h or more before the RD onset yielded interesting results on the composite structure and spatial distribution differences. Contrary to the expectation from hypothesis A3, and the component from hypothesis A2 relating to AC lifetimes pre-RD, the onset group is less

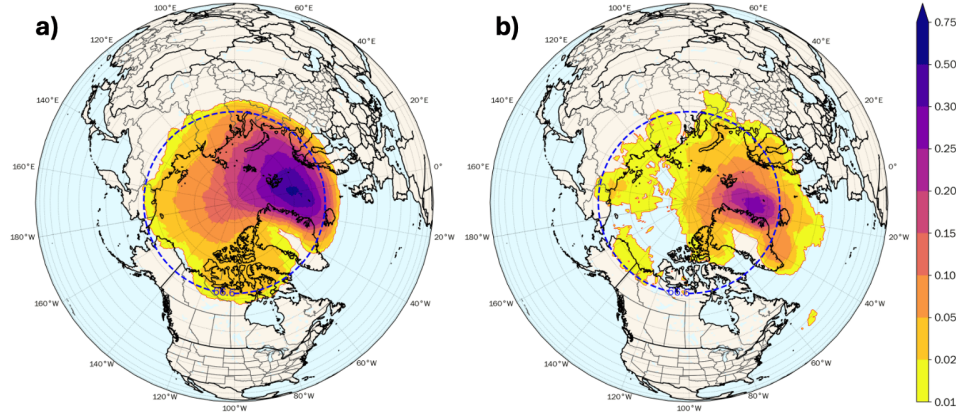


Figure 3.1: Occurrence frequency (%) of (a) all 6 h cyclone positions where RD criteria is met, and (b) cyclogenesis locations of all cyclones that meet RD criteria. Occurrence frequency is calculated as the percent of 6-h time steps between 1979–2022 where a grid point is located within 500 km of the portion of the track of an AC where the criteria specified above is met. The Arctic Circle is denoted by the dotted blue line.

avored in the central Arctic basin, and is associated with a farther away closest TPV at the onset of RD, relative to the 24-hour group.

Finally, a comparison of rapidly deepening ACs and ordinary ACs that intensify but fail to meet RD criteria yields some results that are comparable to midlatitude rapidly deepening cyclone characteristics, such as greater baroclinic instability relative to ordinary cyclones (e.g., Bentley 2018). The addition of TPVs into this analysis adds some complications, however, as there is no significant difference in the proximity of the nearest TPV between rapidly deepening ACs and ordinary ACs. While TPVs do play an important role in AC evolution and rapid deepening, as over 90% of ACs are associated with at least one TPV (Cavallo and Burg 2024), the distance from the nearest TPV may be less important than the intensity, size, and symmetry of the TPV, which can affect the magnitude of differential cyclonic vorticity advection associated with the TPV.

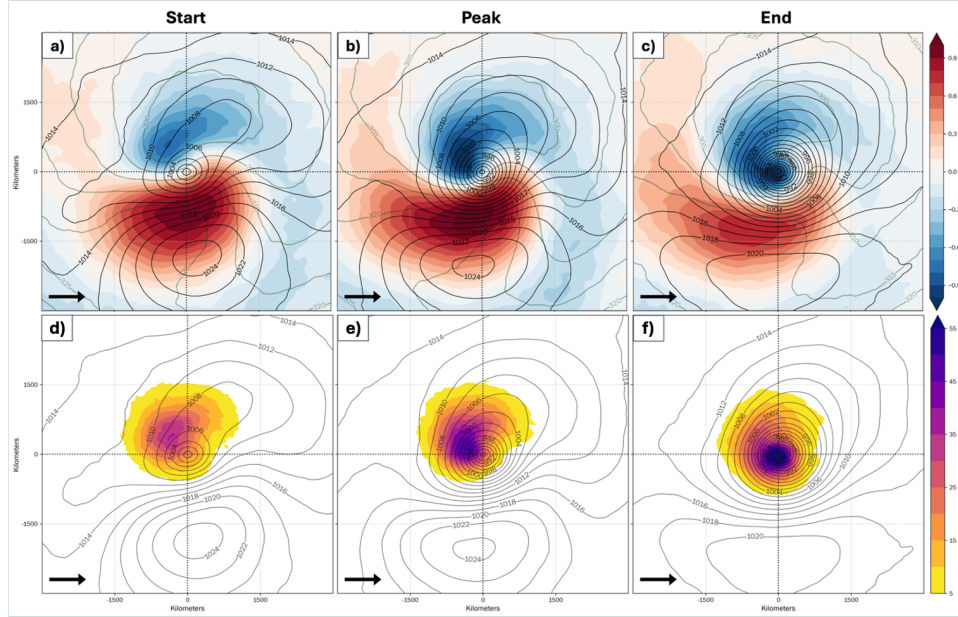


Figure 3.2: Top row shows annual Arctic-wide composite mean of 2 PVU potential temperature (K, green contours), normalized anomaly (sigma, fill), and MSLP (hPa, black contour) at the (a) start, (b) peak, and (c) end of the rapid deepening episode. Bottom row shows the percent of TPV basins in a gridpoint out of all basins for the closest TPV to the AC, at the (d) start, (e) peak, and (f) end of the rapid deepening episode.

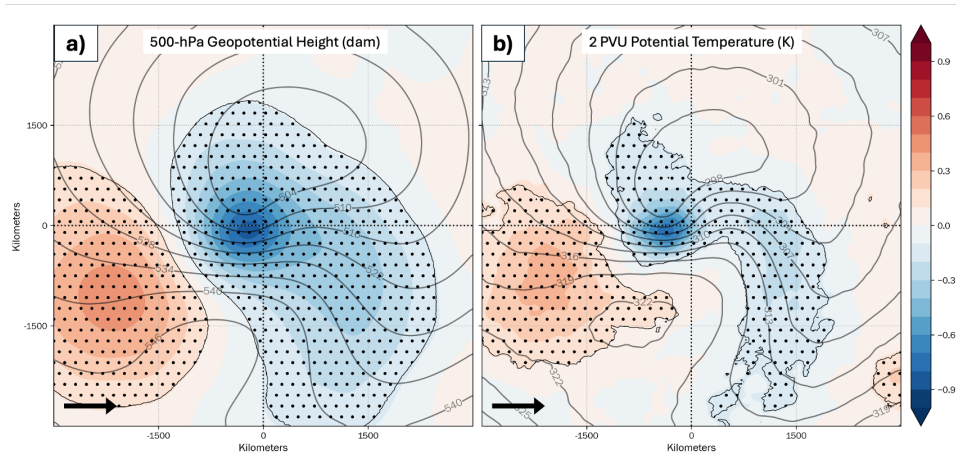


Figure 3.3: Difference in composite mean (a) 500-hPa geopotential height and (b) 2 PVU potential temperature normalized anomalies at the peak minus onset of the rapid deepening episode. Contours show the average of the start and peak composites for (a) 500-hPa geopotential height (dam), and (b) 2 PVU potential temperature (K). Dots denote differences statistically significant at the 99% level.

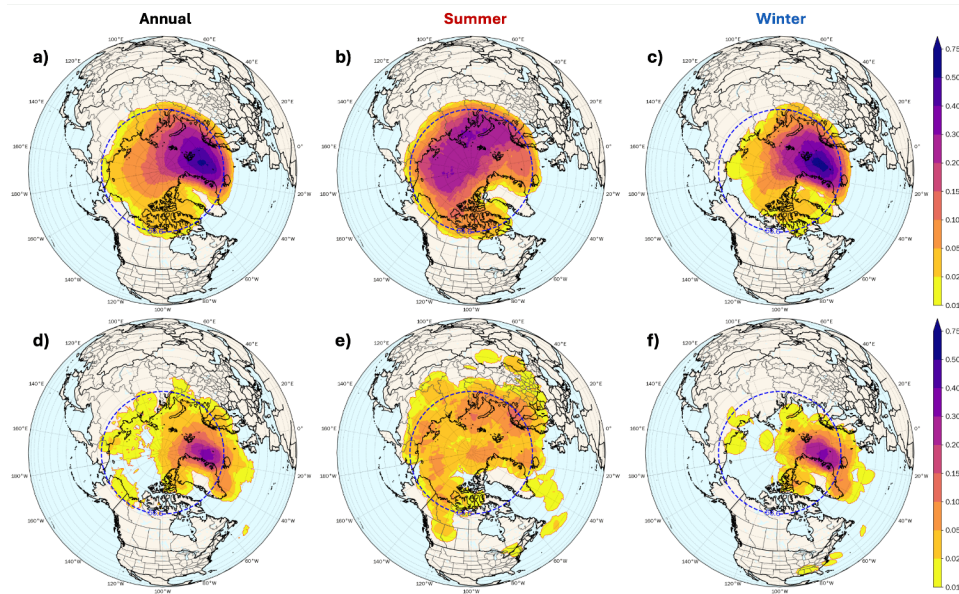


Figure 3.4: Same as Fig. 3.1, but for locations of rapidly deepening ACs for (a) annual cases, (b) summer cases, and (c) winter cases. Bottom row shows cyclogenesis location density for (d) annual cases, (e) summer cases, and (f) winter cases.

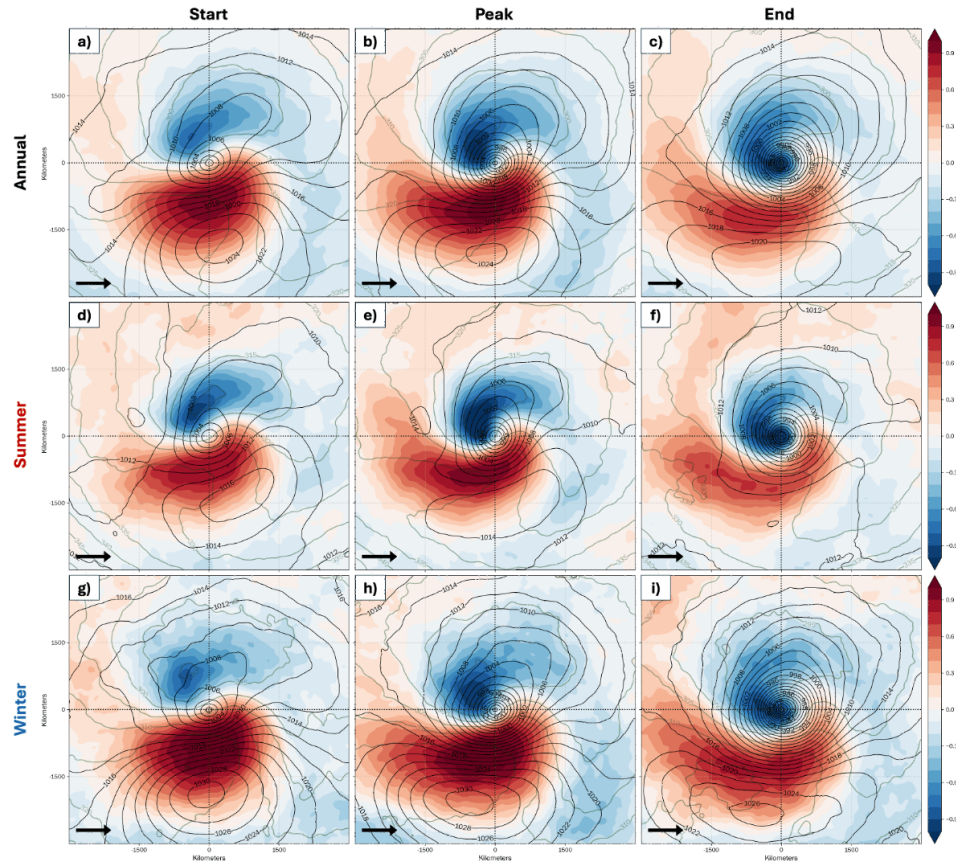


Figure 3.5: Same as top row of Fig. 3.2, for annual cases, showing the (a) start, (b) peak, and (c) end of the RD episode. Second row (d-f) shows the same for summer cases, and the third row (g-i) shows the same for winter cases.

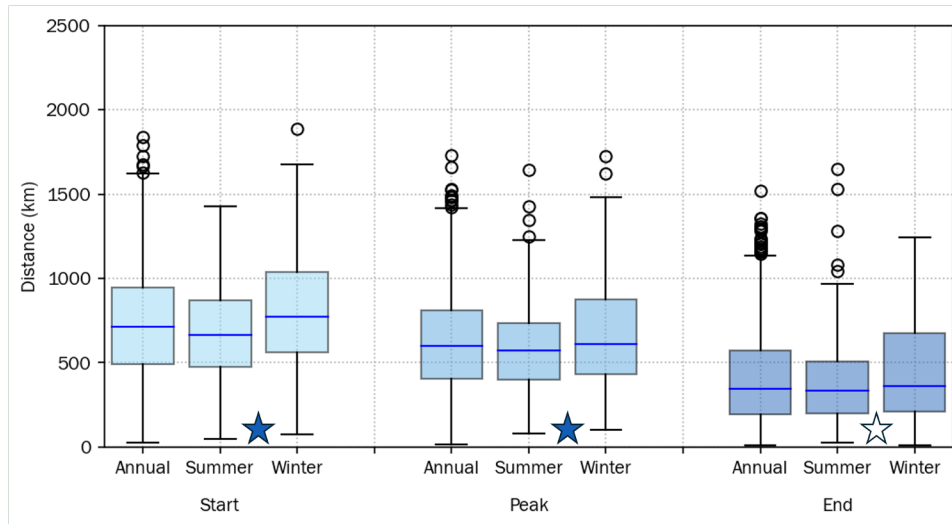


Figure 3.6: Box and whisker plot for the distribution of AC distance from the nearest TPV mass centroid for the start, peak, and end of the RD episode between annual, summer, and winter cases. Blue stars denote differences between the summer and winter distributions are statistically significant at the 99% level, and white stars show statistical significance at the 95% level.

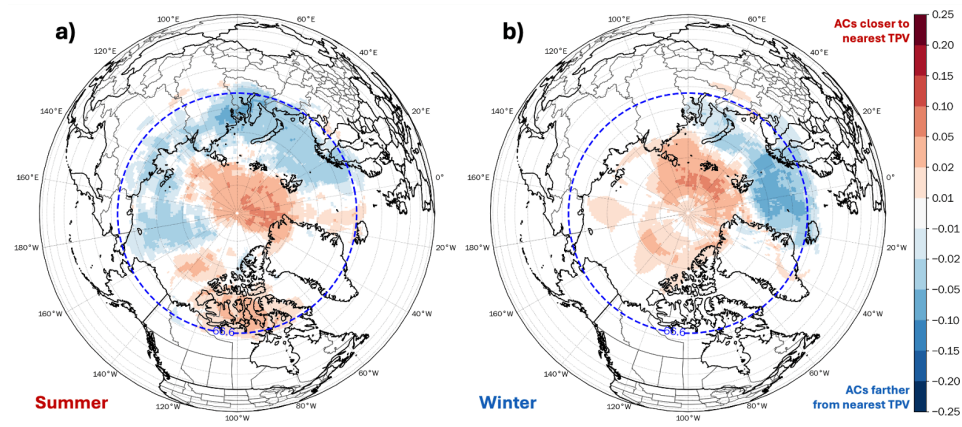


Figure 3.7: Difference in occurrence frequency for cyclone location density at the peak of the RD episode, for cases with an AC-TPV distance below the lower quartile minus cases with an AC-TPV distance above the upper quartile. (a) shows this difference for summer cases, and (b) shows this difference for winter cases.

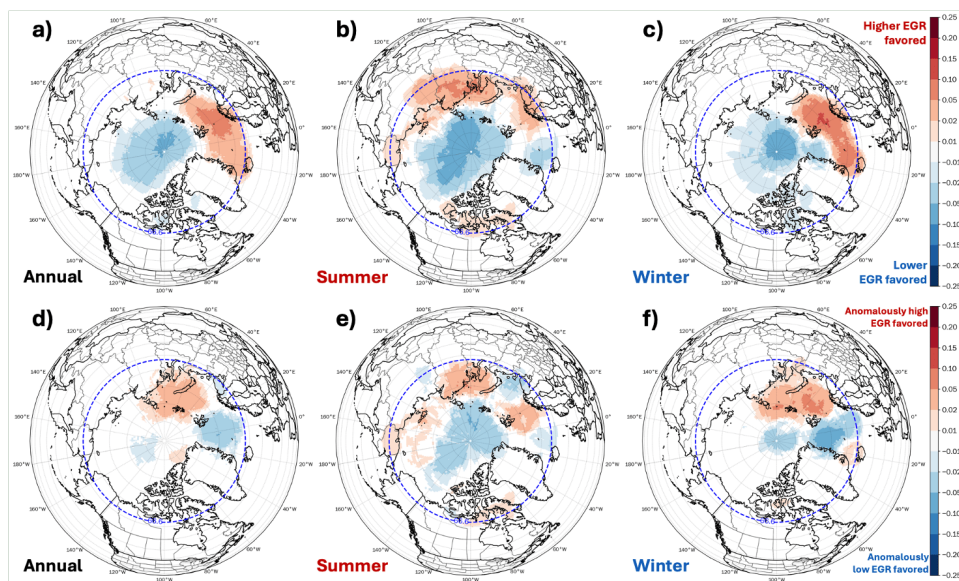


Figure 3.8: Difference in occurrence frequency for cyclone location density at the peak of the RD episode, for (top row) cases with EGR above the upper quartile minus cases with EGR below the lower quartile, for (a) annual cases, (b) summer cases, and (c) winter cases. Bottom row shows the same as the top row, but for EGR normalized anomaly.

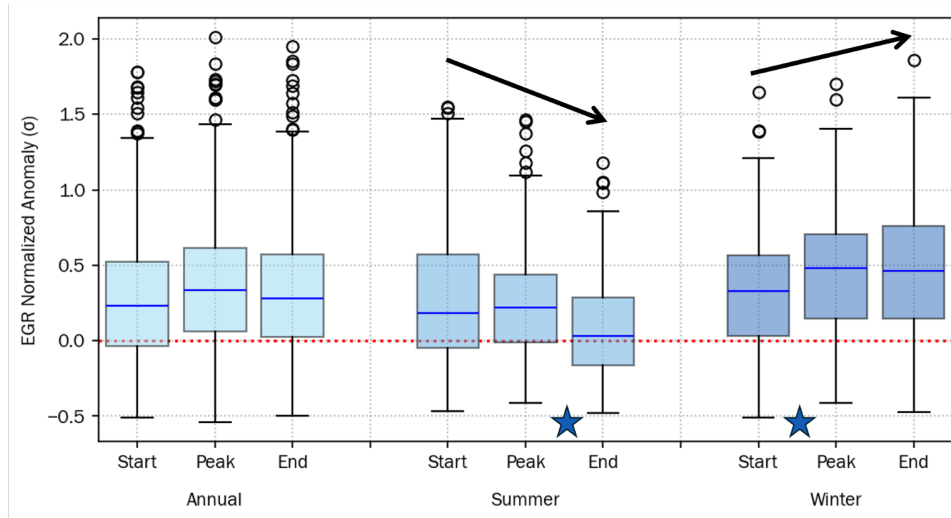


Figure 3.9: Box and whisker plot for the distribution of EGR normalized anomaly for the start, peak, and end of the RD episode between annual, summer, and winter cases. Blue stars denote differences between the start and peak, or peak and end, distributions are statistically significant at the 99% level, and white stars show statistical significance at the 95% level. No stars indicate no significant differences. The black arrows highlight the differing trend in composite EGR normalized anomaly, most prominent in the upper quartiles, throughout the RD episode between the summer and winter cases.

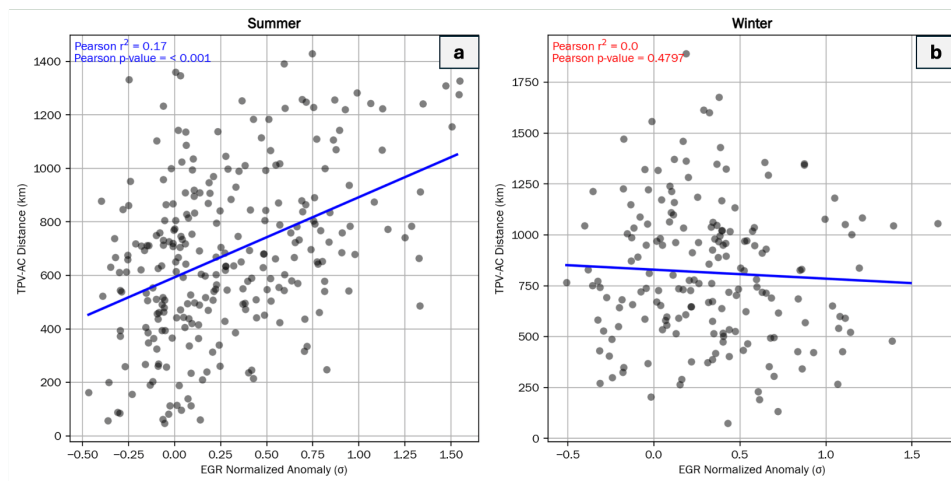


Figure 3.10: Correlation between EGR normalized anomaly and distance from the nearest TPV for (a) summer cases, and (b) winter cases. The Pearson r^2 value and p-value are displayed in the upper left corner, with blue denoting the correlation is statistically significant.

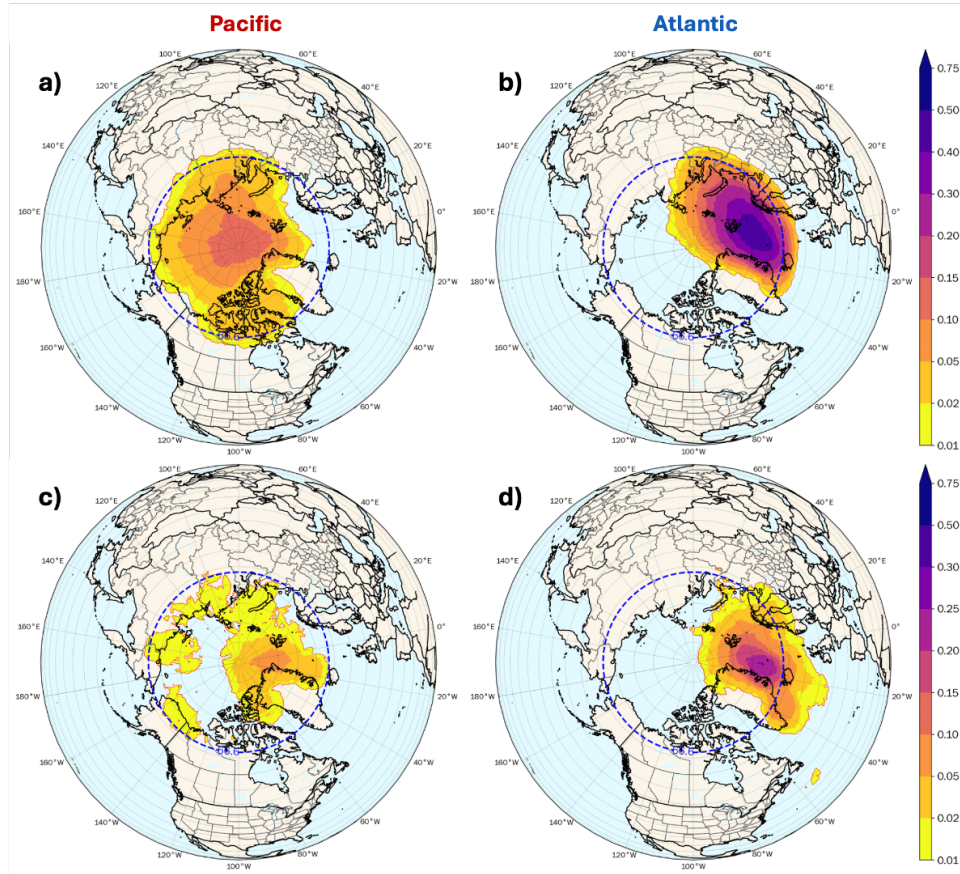


Figure 3.11: Same as Fig. 3.4, but for top row showing RD density for (a) Pacific cases and (b) Atlantic cases, and bottom row showing cyclogenesis location density for (c) Atlantic cases, and (d) Pacific cases.

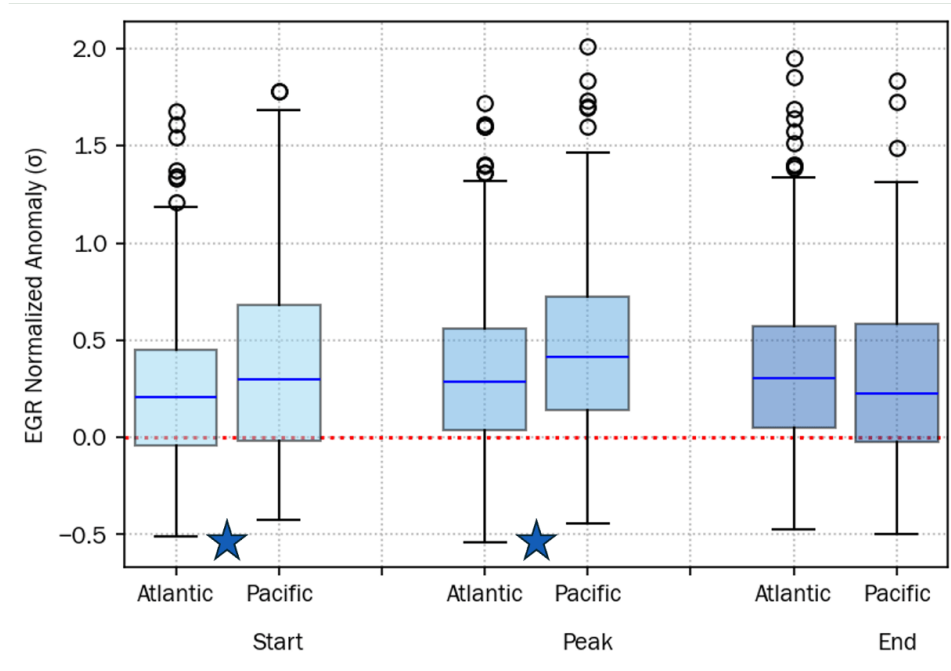


Figure 3.12: Box and whisker plot for the distribution in EGR normalized anomaly for the start, peak, and end of the RD episode between Atlantic and Pacific cases. Blue stars denote differences between Atlantic and Pacific cases are statistically significant at the 99% level, and white stars show statistical significance at the 95% level. No stars indicate no significant differences.

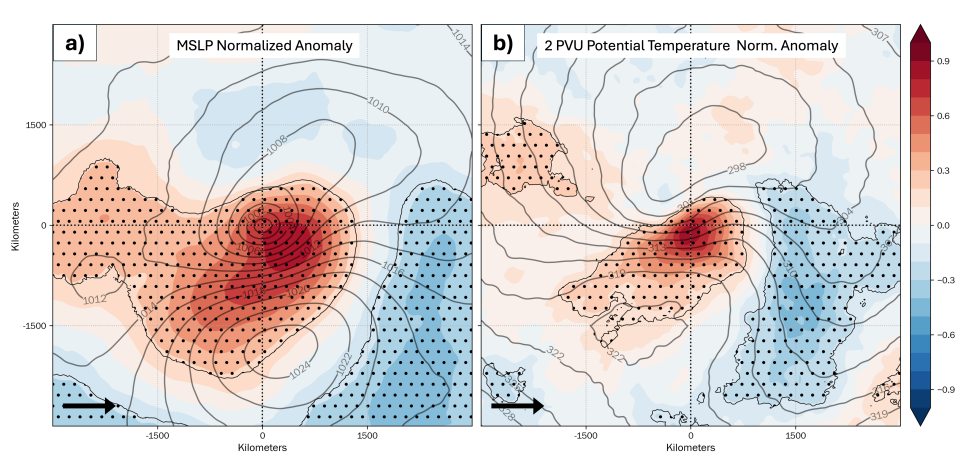


Figure 3.13: Difference in composite mean (a) MSLP and (b) 2 PVU potential temperature normalized anomalies for cases whose RD coincides with cyclogenesis minus cases whose RD onset occurs at least 24 h or longer after cyclogenesis. Contours show the average of the start and peak composites for (a) MSLP (dam), and (b) 2 PVU potential temperature (K). Dots denote differences statistically significant at the 99% level.

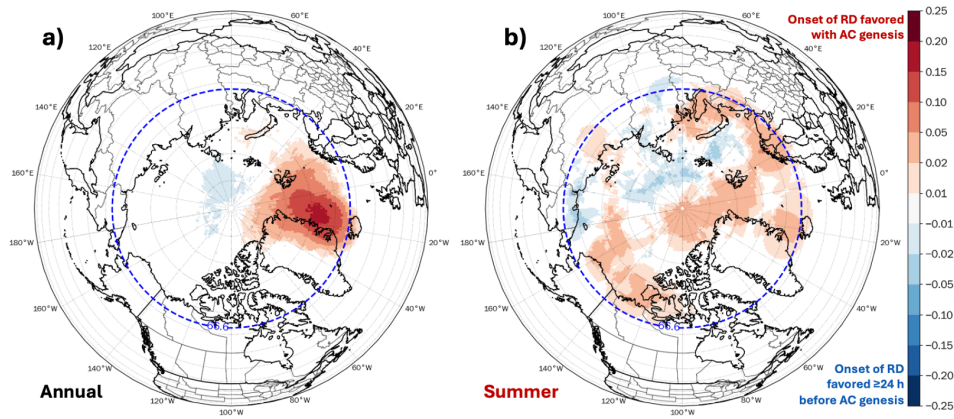


Figure 3.14: Occurrence frequency for cyclone location density at the onset of the RD episode for cases whose RD onset coincides with cyclogenesis minus cases whose RD onset occurs at least 24 h or longer after cyclogenesis, for (a) annual cases, and (b) summer cases. Red shading denotes locations where cyclones in the onset category are more frequent, and blue shading denotes locations where cyclones in the 24 hour category are more frequent.

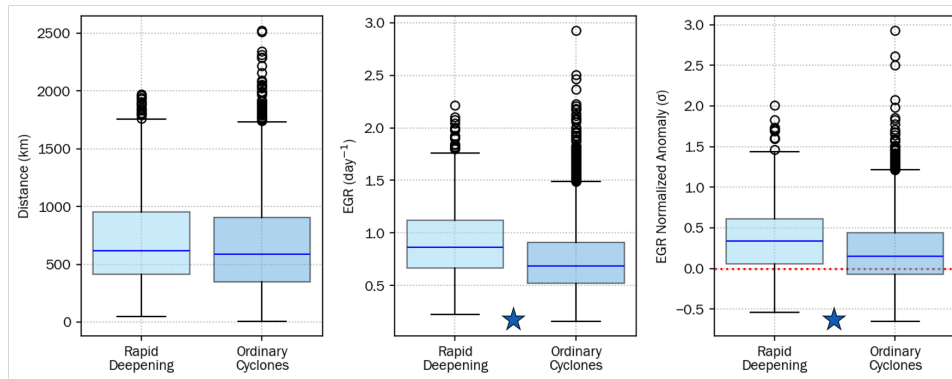
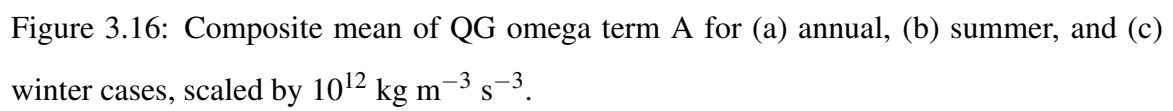


Figure 3.15: Box and whisker plot comparing distributions between rapidly deepening and ordinary ACs for (a) distance from the nearest TPV, (b) EGR, and (c) EGR normalized anomaly. Blue stars denote differences between Atlantic and Pacific cases are statistically significant at the 99% level, and white stars show statistical significance at the 95% level. No stars indicate no significant differences.



Chapter 4

TPV linkages to polar low and Arctic cyclone cases

Chapter 3 presented a systematic evaluation of the characteristics of rapidly deepening ACs. While systematic analyses are useful for reviewing bulk characteristics of many cyclones, they come at the expense of investigating more nuanced variability within the case list and can obscure unique insights that can be gained from more detailed analyses of outlier or high impact cases. Given these constraints of systematic evaluations, this chapter aims to investigate the role of TPV intensity in the track and intensity of two high-latitude rapidly deepening cyclones from a numerical modeling framework. Specifically, the objective is to determine if modifying the intensity of the TPV has a significant impact on the intensification rate, peak intensity, and track of the surface cyclone, and whether this impact is similar between an AC case study or a PL case study.

The August 2012 Great Arctic Cyclone, introduced in section 1.5.1, is a well known example of a rapidly deepening intense AC. Its peak deepening rate ranks at the 29th percentile of annual RD cases, but increases to the 72nd percentile relative to all summer RD cases. Its association with two TPVs during its rapid deepening phase make it of interest to study numerically to isolate the contribution of each TPV to the deepening rate and track of AC12. A secondary objective of this chapter is to compare the impact of TPV intensity on cyclone intensification between an AC and a PL case by examining an intense February 1987 PL that affected Norway.

4.1 August 2012 great Arctic cyclone

This section reviews the synoptic evolution of AC12, differences between the MPAS control simulation and ERA5 reanalysis, then evaluates the impact of modifying TPV 1, TPV 2, and both TPVs, on the track and intensity of AC12. ERA5 maps accompanying this

section show 2 PVU potential temperature in fill, and MSLP below 1008 hPa contoured every 4 hPa, from the genesis of AC12 leading up to its peak intensity.

4.1.1 ERA5 and control simulation overview

AC12 developed as a weak baroclinic wave over central Russia around 0000 UTC 3 August 2012 (Fig. 4.1a), developing downshear of TPV 1. Gradual deepening occurred over the next two days (Fig. 4.1b-c) as AC12 tracked northeastward towards the northern Siberian coast. Contributions to its intensification at this stage primarily came from differential cyclonic vorticity advection downstream of TPV 1 and its associated synoptic-scale trough (not shown). Around the same time, an antecedent decaying AC and its associated TPV 2 meandered over the Arctic Ocean north of Siberia.

As AC12 moved offshore into the Arctic Ocean on 5 August 2012 (Fig. 4.1d), it began to undergo a more substantial RD episode. During this deepening phase, contribution to forcing for ascent from thermal advection increased in addition to contribution from differential cyclonic vorticity advection downstream of TPV 1 (not shown), while TPV 2 looped to the southeast, becoming situated upstream of TPV 1. By 6 August 2012 (Fig. 4.1e-f), AC12 reached its peak intensity of about 962 hPa as it came into vertical alignment with TPV 1 and stalled, while the antecedent AC previously associated with TPV 2 merged into the large circulation of AC12. The stalling of AC12 at its peak intensity over a MIZ contributed greatly to an associated rapid sea ice loss event (Zhang et al. 2013).

The control simulation was initialized at 0000 UTC 4 August 2012 and ran through 0000 UTC 11 August 2012. The broad synoptic evolution of AC12, its associated TPV 1, and antecedent TPV 2 and its associated AC in the control simulation (Fig. 4.2) is largely similar to ERA5 reanalysis, as it captures the rapid deepening of AC12 downstream of TPV 1, TPV 2 looping behind TPV 1, and the vertical coupling and stalling of AC12 and TPV 1. The main differences between the control simulation and ERA5 are in the sub-synoptic scale, as the RD episode is slightly delayed and more rapid in the MPAS control simulation

than ERA5 (Fig. 4.3). In the control simulation, AC12 reaches a peak intensity about 3 hPa deeper, and about 3–6 h later, relative to ERA5. Despite these minor differences, this lends confidence to the control simulation adequately representing the observed evolution of AC12 and associated TPVs.

4.1.2 TPV modification simulations

The first experiment consists of several simulations intended to modify the intensity of TPV 1, with an additional simulation designed to weaken the broader upper-tropospheric trough that TPV 1 is embedded within. The model setup and naming convention of these simulations are described in more detail in section 2.3.2. The result of applying an upper-tropospheric cooling tendency to intensify the TPV is shown in Fig. 4.4, which compares the vertical structure of the TPV between the control simulation and the TPV1++ simulation at forecast hour 24. Both simulations show a classic TPV cross section profile, with a depressed tropopause and a locally maximized EPV anomaly relative to the background environment centered at the vortex core. After applying an upper-tropospheric cooling tendency for a 24 h window (Fig. 4.4b), the resulting TPV has a lower tropopause level, larger EPV magnitude in the vortex core, and a larger spatial extent of the TPV, relative to the TPV in the control simulation. This indicates the resulting TPV is more intense and associated with a stronger cold pool relative to the control simulation.

The remainder of this section discusses the impacts of intensifying TPV 1, TPV 2, both TPVs, and weakening the trough TPV 1 is embedded in. For conciseness and due to close similarities in the key results between the TPV1+ and TPV1++, TPV2+ and TPV2++, and Both+ and Both++ simulations, only the TPV1++, TPV2++, Both++, and Trough-simulations, will be discussed.

The key differences between the control simulation and the TPV1++ simulation are summarized in Fig. 4.5. Both simulations begin with near identical surface cyclone track

and intensity, but begin to diverge noticeably after 0000 UTC 5 August 2012 as AC12 begins to rapidly deepen. During the rapid deepening phase of AC12, the TPV1++ simulation has slightly less positive 700-hPa geostrophic warm air advection over the center of AC12 than the control simulation. This is a result of both a slower and farther left cyclone track, and a more expansive region of cold air advection due to the stronger cold pool associated with TPV1. Differential geostrophic cyclonic vorticity advection over the center of AC12, however, is much greater in the TPV1++ simulation than the control simulation, and is associated with a greater deepening rate of AC12 relative to the control simulation on 5 August 2012. After 0600 UTC 6 August 2012, the TPV1++ simulation has less positive differential geostrophic vorticity advection, a slower MSLP deepening rate, and a more rightward shifted track of AC12, relative to the control simulation. This is indicative of the stronger TPV resulting in AC12 maturing and peaking in intensity about 3–6 h earlier in the TPV1++ simulation than the control simulation. The minimum MSLP of AC12 in the TPV1++ simulation is 0.75 hPa lower than the control simulation.

Repeating the analysis above for the TPV2++ simulation (Fig. 4.6), there is more positive differential geostrophic vorticity advection over the center of AC12 in the TPV2++ simulation relative to the control simulation by 24-h forecast lead time. This is due to lower geopotential heights associated with TPV2 situated upstream of AC12, and a more meridional orientation of TPV2, relative to the control simulation (not shown). The increase in differential geostrophic vorticity advection relative to the control simulation is less than that in the TPV1++ simulation, however, and the minimum MSLP of AC12 in the TPV2++ simulation is only 0.32 hPa lower than the control simulation.

As TPV 1 was embedded within a broad upper-tropospheric trough, the Trough- simulation is used to compare the impacts of modifying the smaller-scale TPV1 relative to modifying the larger-scale synoptic trough. The most pronounced impact by forecast hour 48 is a weaker magnitude and more positively tilted 500-hPa trough, and a more progressive embedded TPV, relative to the control simulation (Fig. 4.7). A weaker trough results

in weaker differential cyclonic vorticity advection, and the spatial peak magnitude of 500-hPa geostrophic absolute vorticity advection during the peak deepening rate of AC12 is substantially weaker relative to the control simulation (Fig. 4.7).

In contrast to intensifying TPV1 and TPV2, the Trough- simulation has greater 700-hPa warm air advection relative to the control simulation over the center of AC12 (Fig. 4.8). This is due to a more progressive surface cyclone expanding farther from the region of cold air advection associated with the cold pool of TPV1, despite overall weaker peak warm air advection magnitude downstream of the trough relative to the control simulation due to a weaker cyclone and associated circulation infringing upon its embedded baroclinic zone. At the same time, differential cyclonic vorticity advection during the peak deepening rate of AC12 is much less positive, with a resulting decreased peak deepening rate of AC12, relative to the control simulation.

An overview of the TPV1++, TPV2++, Both++, and Trough- simulations is shown for minimum MSLP and 12-h MSLP change (Fig. 4.9), and surface cyclone tracks (Fig. 4.10). Between the TPV1++ and TPV2++ simulations, intensifying TPV1 has a greater role in intensifying AC12 than intensifying TPV2, as the peak deepening rate of AC12 is shifted earlier and with a prolonged duration relative to the control simulation, resulting in a stronger and earlier peak intensity. The combined impact of intensifying both TPVs results in the greatest intensity and deepening rate of AC12 relative to intensifying individual TPVs, with the greatest contribution from TPV1.

Intensifying TPV2 leads to only a marginally stronger peak intensity of AC12. Still, it has a greater impact on the track of AC12 relative to TPV1, as the more meridional orientation of TPV2 results in a slight leftward shift in the synoptic-scale flow and steers AC12 farther to the left once it is situated downstream of TPV2. By forecast hour 72 valid 0000 UTC 7 August 2012, at the end of its deepening phase, the TPV1++ simulation situates AC12 26 km northeast of the control simulation, with the eastward shift in position due to the earlier occlusion of the cyclone, and the TPV2++ simulation situates AC12 40

km southwest of the control simulation. The combined impact of intensifying both TPVs situates AC12 16 km east of the control simulation.

Weakening the synoptic-scale trough TPV1 is embedded within results in larger magnitude shifts in both track and intensity of AC12 relative to intensifying either or both TPVs. The weaker and more positively tilted trough results in a more progressive and rightward shifted track in AC12 relative to the control simulation. Differences in the Trough- and control simulations become most pronounced during the peak intensity of AC12, as the cyclone fails to turn left and complete a cyclonic loop in the Trough- simulation. The peak intensity of AC12 is 4.2 hPa weaker than the control simulation, with the position of AC12 valid 0000 UTC 7 August 2012 situated 217 km northeast of the control simulation.

4.2 February 1987 polar low

This section reviews the synoptic evolution of PL87, differences between both control simulations performed, and the IC uncertainty ensembles performed for both control simulation initializations. ERA5 maps accompanying this section show 2 PVU potential temperature in fill and MSLP contoured every 4 hPa.

4.2.1 ERA5 and control simulations overview

The precursor low to PL87 developed as an inverted surface trough south of Svalbard around 0000 UTC 25 February 1987, in association with lee cyclogenesis downwind of Svalbard initially characterized by an inverted surface trough (Fig. 4.11a). At the same time, a strong and relatively symmetric TPV exited the Arctic Ocean and was moving south towards Svalbard, situated upstream of PL87, while a separate filament of high PV was moving east off the coast of Greenland. Over the next day (Fig. 4.11b-c), the TPV became increasingly deformed while interacting with the PV filament exiting Greenland. More substantial surface cyclogenesis occurred between 1200 UTC 25 February 1987 and 0000 UTC 26 February 1987 as PL87 was situated downstream of both upper-tropospheric

PV anomalies. Throughout 26 February 1987 (Fig. 4.11d-e), PL87 came into vertical alignment with the TPV. It developed an increasingly axisymmetric structure with an associated warm core, denoting its transition from a primarily baroclinic cyclone to a warm-core cyclone associated with shallow convection. PL87 subsequently tracked southwards, making landfall in Norway early on 27 February 1987 and rapidly dissipating inland thereafter. This transition also resembles that from an initial forward-shear structure during the cyclogenesis phase of PL87 to a reverse-shear structure following the warm seclusion of PL87 (Stoll et al. 2021).

Two control simulations were performed for this experiment. The first control simulation was initialized on 0600 UTC 25 February 1987, with the second simulation initialized on 1200 UTC 25 February 1987. There is a stark difference in forecast surface cyclone track and intensity between both simulations (Fig. 4.12). The first control simulation fails to develop a surface cyclone that undergoes warm seclusion and transition into an axisymmetric barotropic cyclone, instead depicting a weak and progressive surface cyclone rapidly tracking eastward towards Novaya Zemlya. This represents a remarkable failure to simulate an observed cyclone evolution in as little as 12–18 h forecast lead time. In contrast, the second control simulation shows an evolution similar to the analyzed ERA5 evolution, with PL87 rapidly deepening, occluding, and transitioning into an axisymmetric warm-core low before moving south into Norway.

There is a clear inflection point before which simulations do not show PL development, and after which simulations do show notable PL development. While not shown here, simulations initialized before 0600 UTC 25 February 1987 show a similar evolution to the first control simulation, and simulations initialized after 1200 UTC 25 February 1987 show a similar evolution to the second control simulation. As all of these simulations share an identical configuration, the likely source of this inflection point is the ERA5 ICs, more specifically, an update due to data assimilation between 0900 UTC and 1200 UTC 25 February 1987.

4.2.2 IC uncertainty ensemble

While there is value in using deterministic ERA5 data for ICs, in part due to its relatively high temporal and spatial resolution, high-latitude in-situ observations are sparser in the Arctic than in lower latitudes, and satellite quality has improved in recent decades. Accordingly, uncertainty in ERA5 ICs is greater in earlier years than in later years of the dataset. As ERA5 provides a 10-member IC uncertainty ensemble, a proposed hypothesis is that if the IC uncertainty ensemble is sufficiently dispersive, at least one member before the inflection point will show the observed development of PL87, and at least one member after the inflection point will not show the observed development of PL87.

Two sets of 10-member ERA5 IC uncertainty ensembles were run, one initialized at 0600 UTC 25 February 1987, and one initialized at 1200 UTC 25 February 1987. The surface cyclone track density from the first ensemble (Fig. 4.13a) shows that no members simulated the observed evolution of PL87. There is variability in the intensity and forward speed of the cyclone, although synoptic spread in the cyclone evolution is limited. The same is true of the second ensemble (Fig. 4.13b), as all members simulated the observed warm seclusion of PL87 and transition to a warm-core axisymmetric cyclone, despite some small-scale variability in the intensity and forward progression of PL87 leading up to its warm seclusion.

The sensitivity of the evolution of PL87 to upstream features is quantified using ESA. Figure 4.14a shows the leading mode of variability in the 0600 UTC 25 February 1987 ensemble MSLP field, represented by EOF1, valid at 1200 UTC 26 February 1987. EOF1 is characterized by a dipole pattern oriented west to east along the modeled track of PL87, with the ensemble mean cyclone center situated slightly within the negative EOF1 region. This indicates that the primary mode of variability is dominated by forward speed variability with a slight component of intensity variability, such that ensemble spread encompasses members that are faster and slightly weaker with PL87, or slower and slightly stronger with PL87.

Using EOF1 of the MSLP field as the response function J , the sensitivity of EOF1 of the MSLP field at 1200 UTC 26 February 1987 to 2 PVU potential temperature is calculated at forecast hour 3, valid at 0900 UTC 25 February 1987 (Fig. 4.14b). Notably, despite the ensemble ICs containing non-negligible variability in the intensity of the TPV (not shown), there is no sensitivity to its intensity. Rather, the greatest sensitivity is in the edge of the vortex and in the amplitude of the ridge upstream of the vortex northwest of Svalbard. Per ESA, a higher amplitude ridge upstream of the TPV is associated with a more intense and slower PL, while a lower amplitude ridge is associated with a weaker and more progressive PL. This indicates that the asymmetry of the TPV, and upstream features that affect its track and symmetry, have a greater role on the evolution of PL87 than the intensity of the TPV. Additionally, the PV filament exiting Greenland is another source of sensitivity, as its orientation influences the location and rate of surface cyclogenesis.

As both ensemble distributions in the track of PL87 fail to overlap with each other, the next objective is to quantify the role of the intensity of the TPV on the track and intensity of PL87. In the previous section of this chapter, a more intense TPV resulted in a more intense and earlier occlusion of AC12. However, this may not necessarily be the case for a PL as the role of TPVs in PL evolution is less conclusive than for ACs.

For the 0600 UTC 25 February 1987 ensemble, a second set of simulations was performed with a stronger TPV to assess how much the distribution is shifted closer to the observed PL87 evolution. Intensifying the TPV resulted in a slight southward nudge in the modeled PL87 track relative to the control ensemble (Fig. 4.15a). While not shown here, this is also accompanied by a slightly slower forward speed of PL87 as well. The impact of a stronger TPV on the intensity of PL87 (Fig. 4.15b) is initially a slightly weaker surface cyclone relative to the control ensemble, later transitioning to a slightly more intense cyclone, although this difference is small and synoptically negligible.

For the 1200 UTC 25 February 1987 ensemble, a second set of simulations was performed with a weaker TPV to assess by how much it shifts the distribution away from the

observed PL87 evolution. Weakening the TPV resulted in a slight eastward displacement in the surface cyclone track (Fig. 4.16a), mostly pronounced after the warm seclusion, in addition to a slightly weaker peak intensity of the surface cyclone (Fig. 4.15b). As with the earlier ensemble before the inflection point, these differences are small and synoptically insignificant.

4.3 Discussion

The results from the AC12 experiment indicate that the intensity and track of AC12 are sensitive to the intensity of both TPVs. The intensity of TPV 1 most strongly influences the intensity of AC12, while the intensity of TPV 2 most strongly influences the track of AC12 and is also associated with a slight variation in the intensity of AC12. The overall impact on TPV intensity on the evolution of AC12 is on the sub-synoptic scale, with variability in the peak minimum MSLP of AC12 on the order of 3 hPa, and on the order of 2.5 hPa per 12 h for the peak deepening rate of AC12. The largest impact on the track and intensity of AC12 resulted from weakening the broad upper-tropospheric trough AC12 is embedded in, which resulted in a much weaker peak deepening rate, weaker cyclonic vorticity advection, and a substantial right and progressive shift in the track of AC12 which resulted in it ultimately failing to stall in the central Arctic Ocean.

The role of both TPVs in the evolution of AC12 was previously studied by Chen et al. (2022), finding that during the slowdown in the intensification of AC12 before its peak deepening rate, forcing for ascent due to cyclonic vorticity advection from TPV 1 was counteracted by cold air advection associated with the cold pool beneath TPV 1. This finding is supported by the control simulation, as during the slowdown in the deepening rate of AC12, it was situated in a region of cold air advection at 700-hPa. As the RD episode was underway, the surface cyclone was situated in a region of 700-hPa warm air advection and Laplacian of 700-hPa temperature advection downstream of AC12, contributing to stronger overall forcing for ascent.

An idealized schematic of the impact of modifying the intensity of TPV1 and the synoptic-scale trough containing TPV1 on the track of AC12 is shown in Fig. 4.17. The main conclusion is that the magnitude of impact on the surface cyclone evolution is dependent on the scale of the upper-tropospheric feature; amplitude uncertainty in a smaller-scale feature such as a TPV will have accordingly smaller scale impact on the track and intensity of a downstream surface cyclone, while amplitude uncertainty in a larger-scale feature such as a synoptic-scale trough will have an accordingly larger impact on the track and intensity of a surface cyclone.

An important limitation is the subjective 2 PVU potential temperature thresholds selected in the methodology to modify the TPV intensity within. While these values were selected carefully to be consistent with the spatial extent and amplitude of each TPV, the results can be sensitive to the criteria selected. For the AC12 case, several simulations were performed using the TPV1+ configuration but with variable radii. Differences in the intensity of AC12 are mostly negligible before the onset of the RD episode, after which differences in the minimum MSLP between these simulations increase to nearly 4 hPa, in addition to slight variability in the timing of AC12 reaching its peak intensity.

A clear result from the PL87 experiment is the strong sensitivity to the ICs. The failure of the MPAS simulations before the inflection point to represent the observed evolution of PL87 in as little as 12–18 h lead time is not unique to this case. There is precedent of poor short-term PL forecast skill, such as a 2011 PL case that was similarly poorly predicted beyond 12 h forecast lead time, even with high-resolution convection-allowing simulations (Doyle and Biernat 2023). It is expected that in earlier years in reanalysis datasets, especially in higher latitudes, uncertainty in the quality of ICs increases. However, even an ensemble initialized from the ERA5 IC uncertainty ensemble fails to simulate the observed evolution of PL87 before the inflection point. It is speculated that the source of this inflection point is related to the spatial coverage and uncertainty of nearby observations and how

they were assimilated into ERA5, and an opportunity for future work is investigating this case from a data assimilation framework.

Overall, this chapter compared the impacts of variable TPV intensity on two different types of high-latitude rapid cyclogenesis. While these are only two case studies, limiting the broader interpretation of their results, an applicable takeaway point is that variable TPV intensity does not equally affect the evolution of all kinds of surface cyclones, and that proximity to a nearby TPV does not necessarily indicate that the intensity of the TPV has a significant impact on the evolution of the cyclone. While completely removing an observed TPV from a simulation may not yield physically realistic results, it can be speculated based on the results of chapters 3 and 4 that synoptic-scale cyclones may still form and intensify in the absence of TPVs, although the presence of a TPV can aid in increasing the intensification rate of a cyclone relative to cases without a TPV. For the PL87 case, despite synoptic maps showing the TPV located directly upstream of PL87, the greatest sensitivity to the forward speed and intensity of PL87 based off the ERA5 IC uncertainty ensemble simulations was not within the vortex core, but rather in the gradient of the vortex and an upstream ridge influencing the symmetry and track of the vortex.

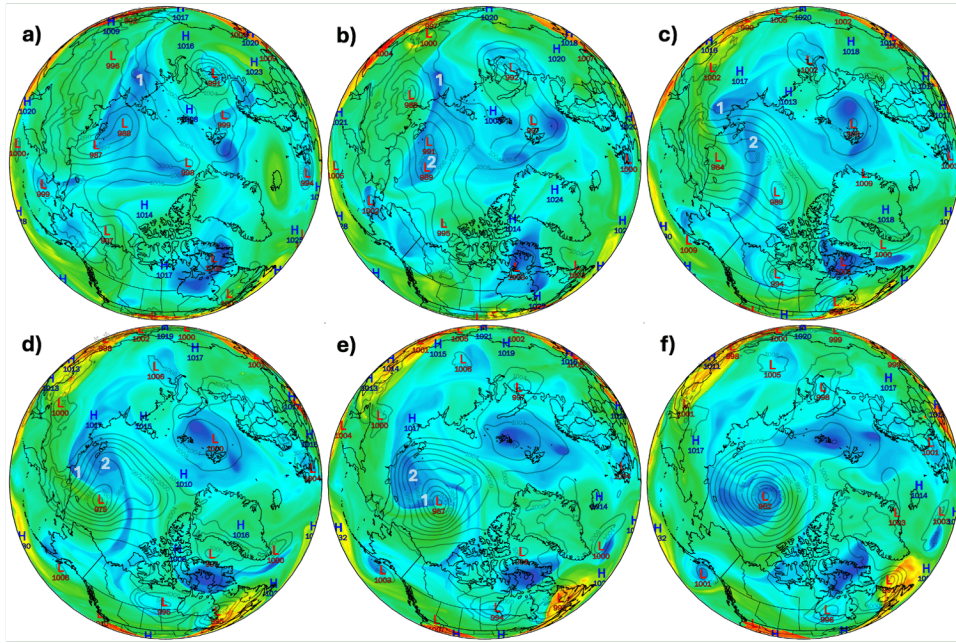


Figure 4.1: ERA5 2 PVU potential temperature (K) in fill and MSLP below 1008 hPa in contours. The locations of TPV 1 and TPV 2 are highlighted where applicable. Frames are valid at (a) 0000 UTC 3 August 2012, (b) 0000 UTC 4 August 2012, (c) 0000 UTC 5 August 2012, (d) 1200 UTC 5 August 2012, (e) 0000 UTC 6 August 2012, and (f) 1200 UTC 6 August 2012.

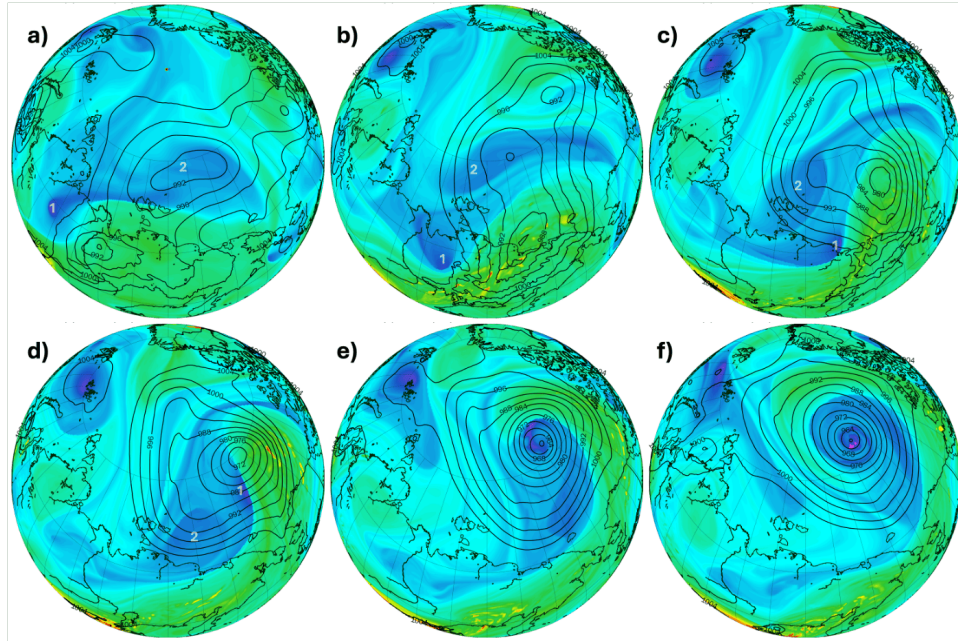


Figure 4.2: MPAS control simulation 2 PVU potential temperature (K) in fill and MSLP below 1008 hPa in contours. The locations of TPV 1 and TPV 2 are highlighted where applicable. Frames are valid at (a) 0000 UTC 4 August 2012, (b) 0000 UTC 5 August 2012, (c) 1200 UTC 5 August 2012, (d) 0000 UTC 6 August 2012, (e) 1200 UTC 6 August 2012, and (f) 0000 UTC 7 August 2012.

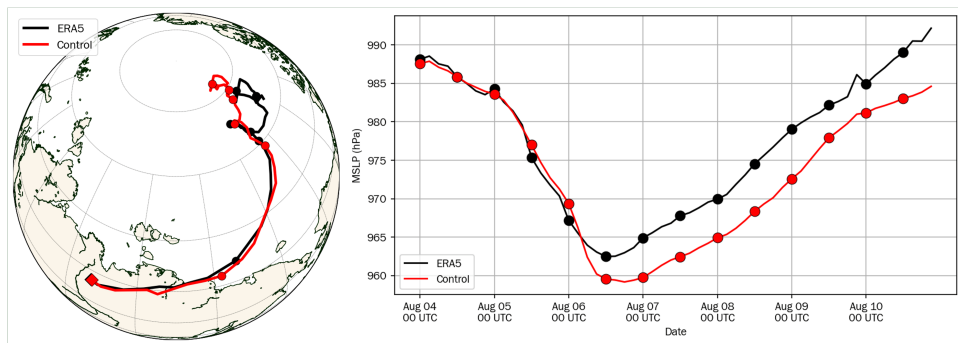


Figure 4.3: Comparison between ERA5 (black) and the control simulation (red) for AC12 cyclone track (left) and minimum MSLP (hPa, right).

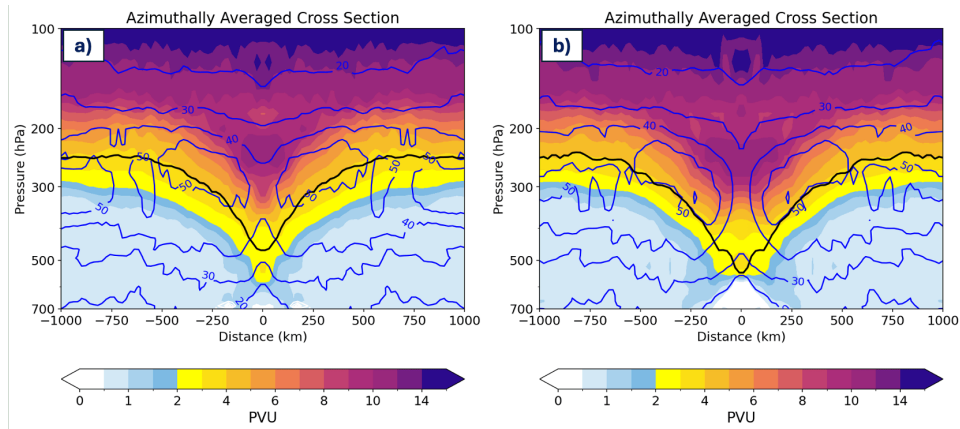


Figure 4.4: MPAS azimuthally averaged cross section centered at the core of TPV 1 at the end of the 24 h TPV modification window, valid 0000 UTC 5 August 2012, for (a) the control simulation, and (b) the TPV1++ simulation. EPV (PVU) is shown in fill, azimuthally averaged wind speed (knots) are in blue contours, and the solid black line denotes the azimuthally averaged tropopause level from the MPAS output.

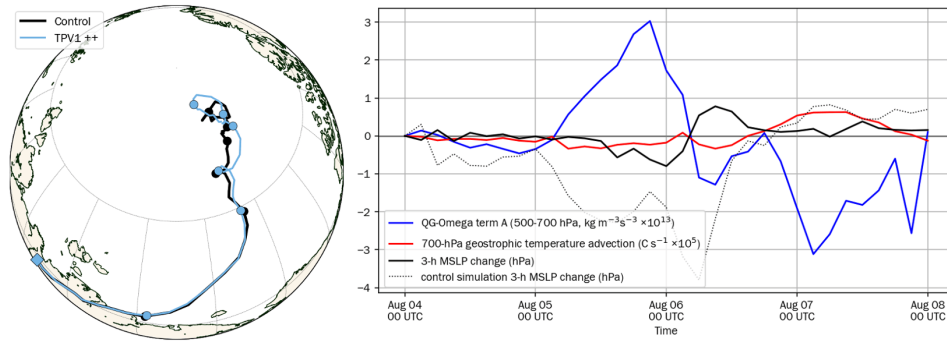


Figure 4.5: (Left) Comparison of AC12 cyclone track between the control simulation (black) and TPV1++ simulation (light blue), with dots every 24 h beginning 0000 UTC 4 August 2012. (Right) Control simulation 3-h MSLP change (hPa, dotted line), difference in the TPV1++ simulation minus the control simulation for term A of the QG omega equation between 500–700 hPa averaged within 4 hPa of the minimum MSLP of AC12 ($10^{13} \text{ kg m}^{-3} \text{ s}^{-3}$, blue line), 700-hPa geostrophic temperature advection averaged within 4 hPa of the minimum MSLP of AC12 (10^5 K s^{-1} , red line), and 3-h MSLP change (hPa, black).

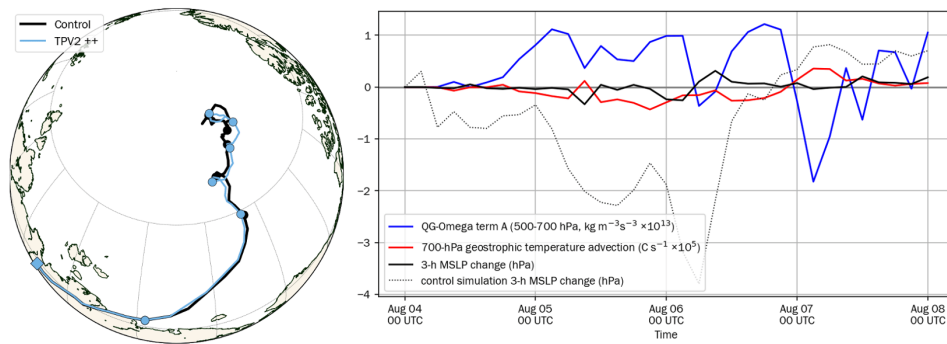


Figure 4.6: Same as Fig. 4.5, but for the TPV2++ simulation.

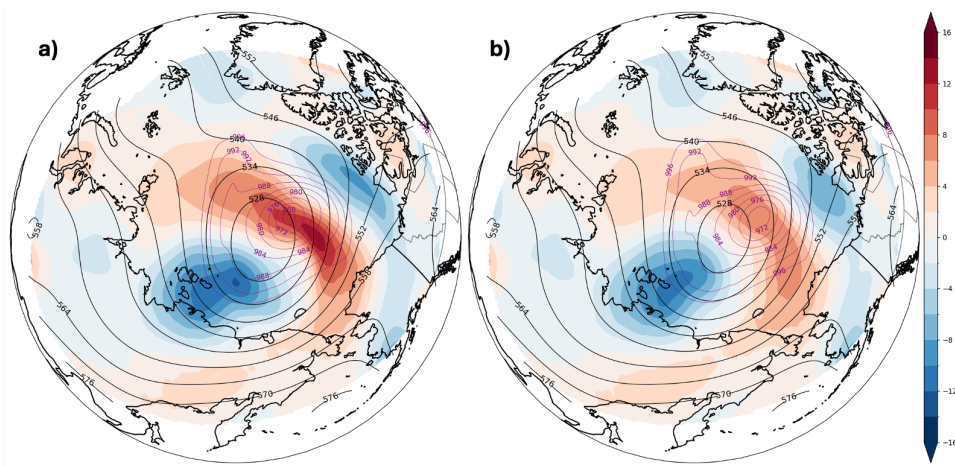


Figure 4.7: 500-hPa geostrophic absolute vorticity advection by the geostrophic wind ($\ast 10^{10} \text{ s}^{-2}$, fill), smoothed 500-hPa geopotential height (dam, black contours), and MSLP below 996 hPa (hPa, purple contours) for (a) control simulation, and (b) Trough- simulation, at forecast hour 51 valid 0300 UTC 6 August 2012.

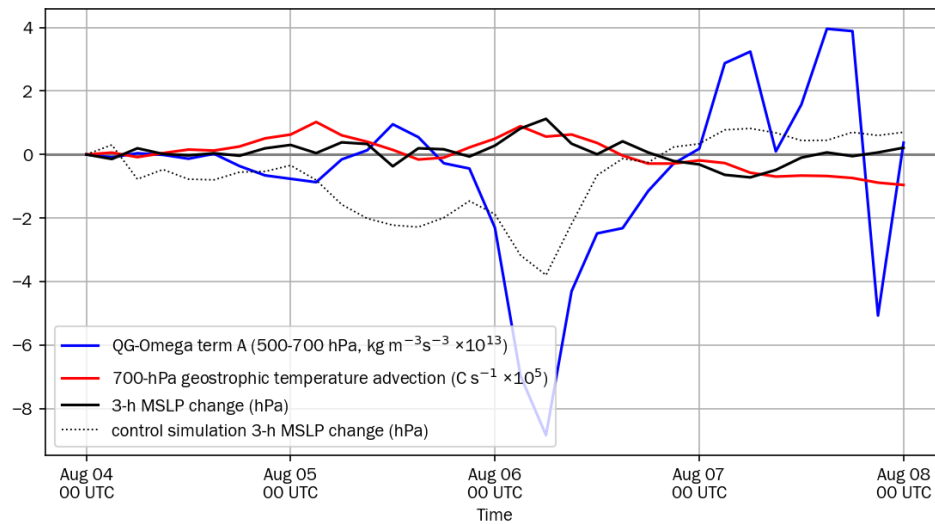


Figure 4.8: Difference in the Trough- simulation minus the control simulation for term A of the QG omega equation between 500–700 hPa averaged within 4 hPa of the minimum MSLP of AC12 ($10^{13} \text{ kg m}^{-3}\text{s}^{-3}$, blue line), 700-hPa geostrophic temperature advection averaged within 4 hPa of the minimum MSLP of AC12 (10^5 K s^{-1} , red line), and 3-h MSLP change (hPa, black).

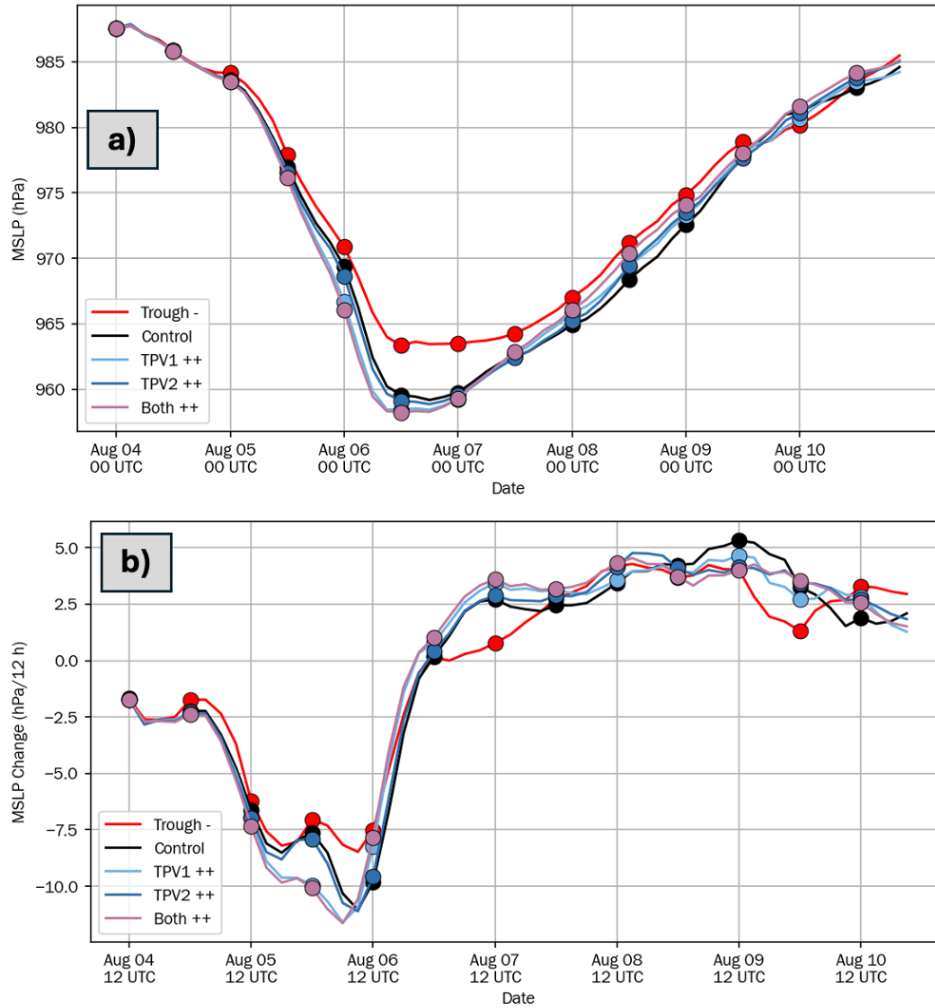


Figure 4.9: (a) AC12 minimum MSLP (hPa), and (b) 12-h change in AC12 minimum MSLP (hPa), for the control simulation (black), TPV1++ simulation (light blue), TPV2++ simulation (dark blue), Both++ simulation (purple), and Trough- simulation (red).

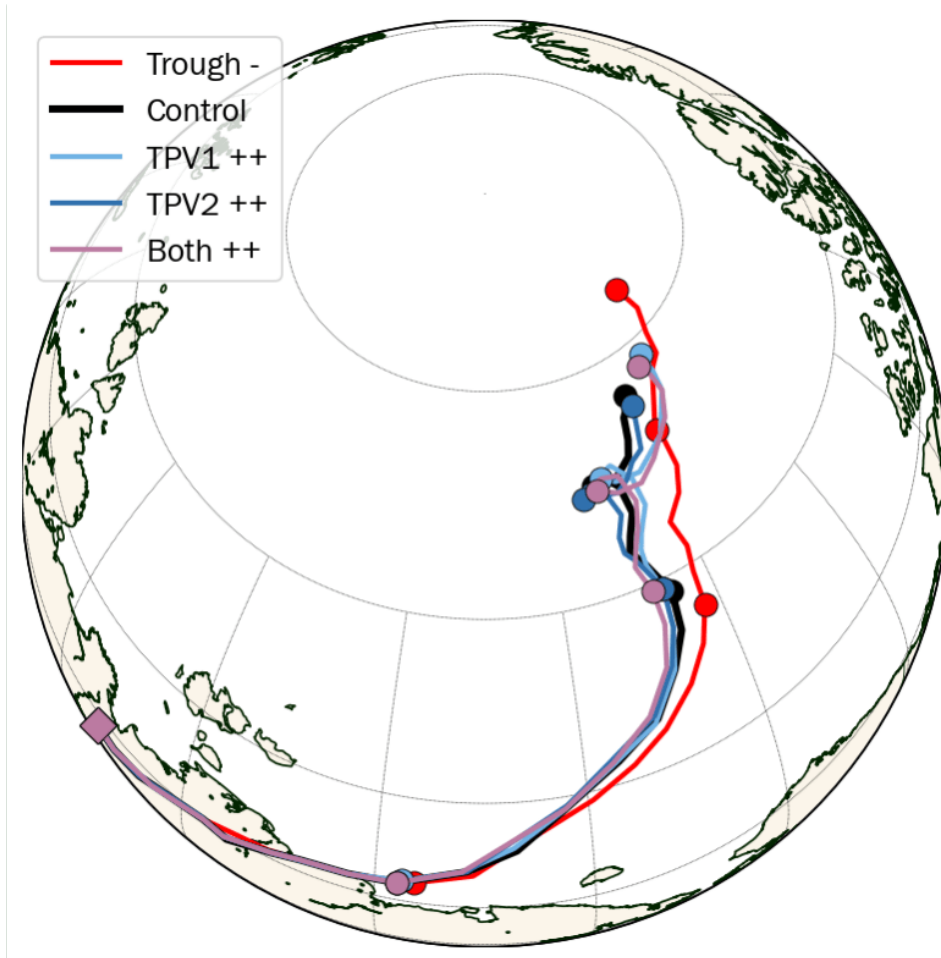


Figure 4.10: AC12 simulated track based on minimum MSLP location for the control simulation (black), TPV1++ simulation (light blue), TPV2++ simulation (dark blue), Both++ simulation (purple), and Trough- simulation (red).

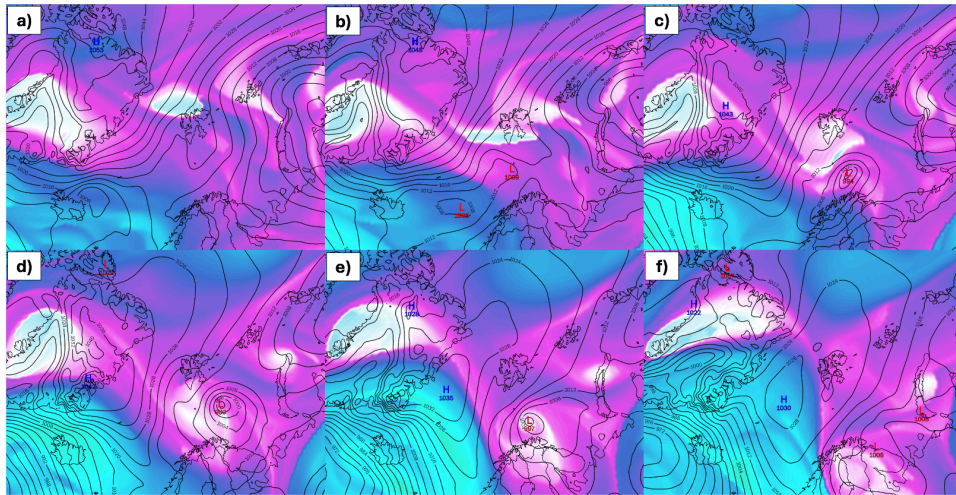


Figure 4.11: ERA5 2 PVU potential temperature (K) in fill and MSLP (hPa) in contours. Frames are valid at (a) 0000 UTC 25 February 1987, (b) 1200 UTC 25 February 1987, (c) 0000 UTC 26 February 1987, (d) 1200 UTC 26 February 1987, (e) 0000 UTC 27 February 1987, and (f) 1200 UTC 27 February 1987.



Figure 4.12: Control simulation cyclone tracks and minimum MSLP (hPa) from MPAS control simulations initialized on 0600 UTC (light blue) and 1200 UTC (dark blue) 25 February 1987. Thicker lines denote lower MSLP, and thinner lines denote higher MSLP.

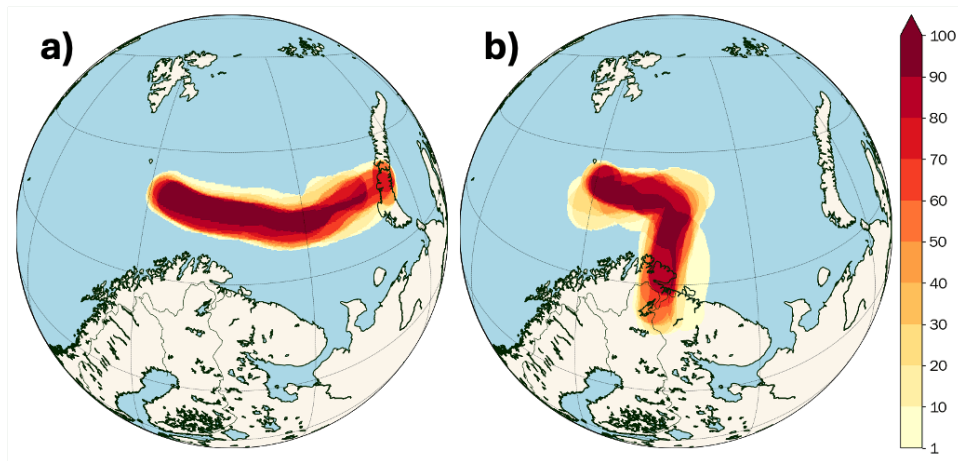


Figure 4.13: Ensemble 75 km track density (%) for the (a) 0600 UTC 25 February initialization, and (b) 1200 UTC 25 February initialization.

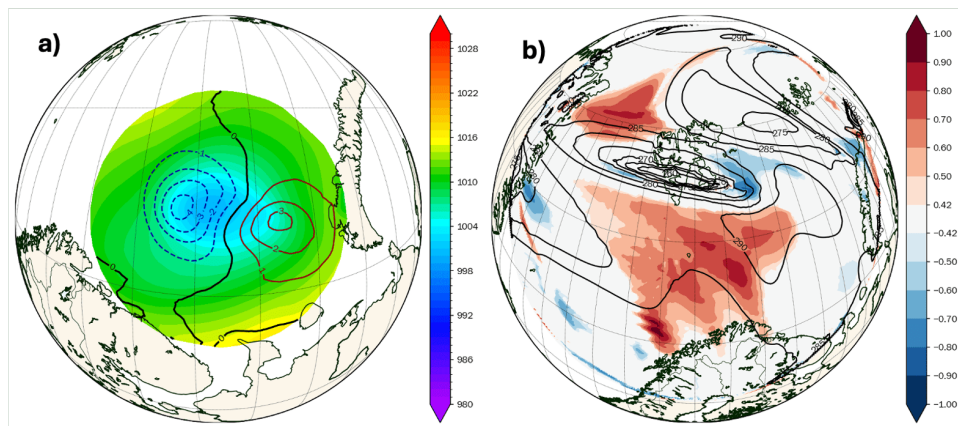


Figure 4.14: ESA analysis for the ensemble initialized 0600 UTC 25 February 1987. (a) Ensemble mean MSLP (hPa) valid 1200 UTC 26 February 1987, with the leading EOF shown in contours. (b) Sensitivity (fill) of the leading EOF in the domain shown in (a), valid at 1200 UTC 26 February 1987, to 2 PVU potential temperature (ensemble mean in contour) valid 0900 UTC 25 February 1987.

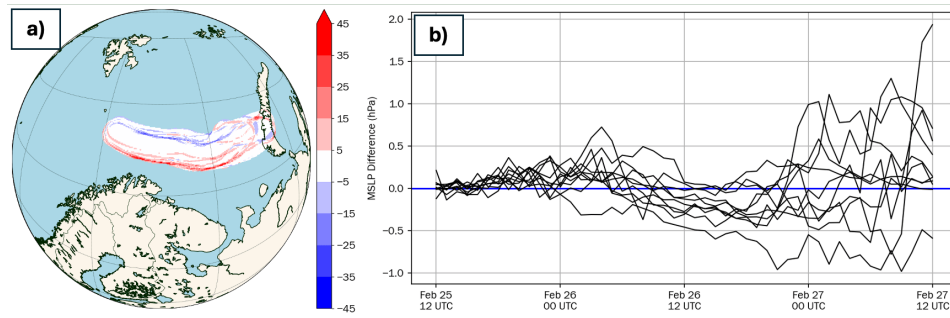


Figure 4.15: (a) 75 km track density difference (%) between the stronger TPV ensemble minus the control ensemble initialized 0600 UTC 25 February 1987. (b) PL87 minimum MSLP difference (hPa) between the stronger TPV ensemble minus the control ensemble.

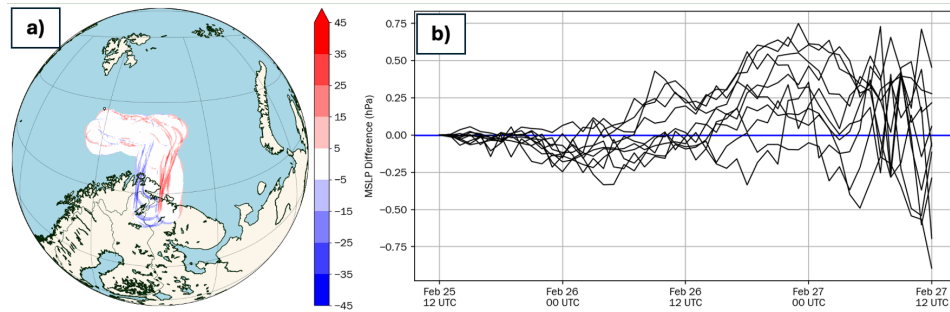


Figure 4.16: (a) 75 km track density difference (%) between the weaker TPV ensemble minus the control ensemble initialized 1200 UTC 25 February 1987. (b) PL87 minimum MSLP difference (hPa) between the weaker TPV ensemble minus the control ensemble.

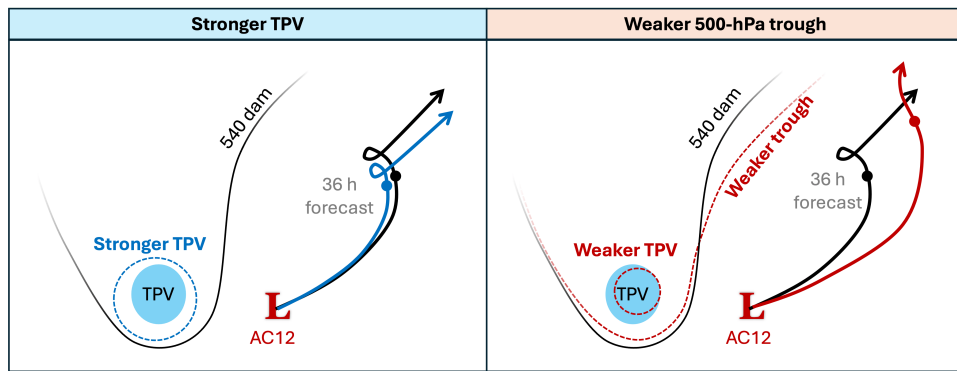


Figure 4.17: Idealized schematic representing the impact of intensifying a TPV embedded within a synoptic-scale trough, versus weakening the amplitude of the synoptic-scale trough, on the track of a developing downstream surface cyclone.

Chapter 5

TPV linkages to the February 2021 cold air outbreak

Chapters 3 and 4 reviewed systematic TPV linkages to rapidly deepening ACs, and individual case studies of TPV linkages to an intense AC and PL, respectively. So far, these analyses have been solely restricted to phenomena within the Arctic. As discussed in Chapter 1, however, TPV impacts are not limited to the Arctic, and TPVs that exit the high latitudes into the midlatitudes can lead to major CAOs (e.g., Papritz et al. 2019; Biernat et al. 2021; Lillo et al. 2021).

Using the TPV track dataset discussed in section 2.1.4 and applying the TPV identification methodology from Szapiro and Cavallo (2018), pathways that TPVs take out of the Arctic and Antarctic into the midlatitudes become most apparent when viewing differences in TPV track density between summer and winter months (Fig. 5.1). In the Northern Hemisphere (Fig. 5.1a), this seasonal difference is asymmetric with more frequent TPVs in winter than summer over northern Canada and eastern Siberia, highlighting these as common TPV pathways out of the Arctic. This aligns with results from Biernat et al. (2021) of common TPV pathways associated with midlatitude CAOs. During the winter, the polar jet stream is stronger and RWs exhibit more frequent intrusions into the Arctic, thus the larger-scale atmospheric flow more readily advects TPVs out of the Arctic into lower latitudes. In the Southern Hemisphere (Fig. 5.1b), this seasonal difference is more symmetric with TPVs more concentrated over the Antarctic during winter than summer. This may be due to the high altitude Antarctic ice sheet inhibiting the frequent RW intrusions into the Antarctic during Austral winter.

This chapter focuses on a case study of a major CAO over North America in February 2021. Minimum temperatures at the peak of the CAO reached -17.8°C (0°F) as far south as central Texas (Fig. 5.2), and this CAO was linked with two TPVs that exited the Arctic

via northern Canada and subsequently merged over southern Canada. The southern Great Plains region in the United States experienced the most severe impacts from this CAO, with major socioeconomic impacts. The presence of multiple merging TPVs adds a unique aspect of this case compared to a similar case study by Lillo et al. (2021) of a CAO linked to a single TPV, which motivates further investigating the role of the multiple TPVs on the CAO evolution in the southern Great Plains from air parcel trajectory, numerical simulation, and ensemble sensitivity frameworks. The goal of this chapter is to better understand the unique impacts of merging TPVs on the evolution of a CAO and its relationship to the intensity of the TPVs, as opposed to a single, intense TPV (e.g., Lillo et al. 2021).

5.1 Case overview

This section reviews the synoptic and mesoscale evolution of the CAO, with a focus on establishing a link between the TPVs and CAO, and providing the necessary background on the case to aid in interpreting the results discussed in section 5.2. ERA5 maps accompanying this section show 2 PVU potential temperature in fill and mean sea level pressure (MSLP) contoured every 4 hPa.

5.1.1 Large-scale precursors

A major large-scale direct precursor to the CAO was an intense cyclone east of the Kamchatka Peninsula in Russia, associated with a negatively tilted upper-tropospheric trough and a downstream ridge (0000 UTC 1 February 2021; Fig. 5.3a). This ridge subsequently expanded poleward and became detached from the midlatitude waveguide (0000 UTC 3 February 2021; Fig. 5.3b), eventually becoming a well-established block, defined as a persistent, quasi-stationary ridge, (e.g., Schwierz et al. 2004) situated north of Siberia over the Arctic Ocean (0000 UTC 5 February 2021; Fig. 5.3c). This evolution led to the displacement of the large-scale tropospheric polar vortex into northern North America, which served as one of several precursors to the eventual CAO.

Over North America, a major winter storm affected the Northeast U.S. in late January associated with a slow-moving trough detached from the polar jet, henceforth referred to as a cutoff trough. As this cutoff trough slowly departed eastward, the zonal progression of the North American waveguide slowed down with the mean jet shifted poleward. Subsequently, a trough and associated cyclone located over the Great Lakes at 0000 UTC 5 February 2021 (Fig. 5.3d) became negatively tilted and contributed to downstream ridge amplification over the Davis Strait and Baffin Bay (0000 UTC 6 February 2021; Fig. 5.3e). This ridge subsequently detached from the waveguide and evolved into a high-latitude block, shifting poleward into northern Canada. In conjunction with the previously described Arctic block, the tropospheric cold pool to the north and associated TPVs were transported into southern Canada along the lee of the Rockies (0000 UTC 6 February 2021; Fig. 5.3f).

From an S2S perspective, the Arctic Ocean and Canada blocks and associated precursor wave breaking events were major forecast barriers to extended range predictability of this event. Subseasonal ensemble members that resolved these blocks at extended lead times were associated with colder temperatures across the southern Great Plains than members that did not adequately resolve both blocks (Millin and Furtado 2022).

5.1.2 TPV and CAO evolution

The poleward expansion of the Canada block discussed in the previous section resulted in a Rex block configuration, with an anticyclone situated poleward of a cutoff trough. Such a configuration tends to slow down the zonal progression of the waveguide. Embedded within the cutoff trough were two TPVs, denoted here as A and B (0000 UTC 8 February 2021; Fig. 5.4a). TPV B subsequently became an elongated PV filament which split into two TPVs, one turning north and west (B1) as it looped cyclonically around TPV A, and another (B2) progressing downstream towards Quebec and Nova Scotia (Fig. 5.4b). TPVs A and B1 rotated about each other over the next several days, and by 13 February merged

into a single TPV (Fig. 5.4b–d). Around the same time, TPV B2 moved east outside of the domain of interest and weakened over the northern Atlantic Ocean.

The first round of the CAO consisted of a shallow cold air damming regime in the lee of the Rockies. During the TPV merger process, a large cold pool associated with the broad upper-level low and maximized beneath both TPVs lingered around southern Canada, situated east of the Canadian Rockies, with 850–hPa temperatures associated with both TPVs ranging between -35°C and -40°C . A strong surface anticyclone situated in the lee of the Rockies amplified by forcing for subsidence upstream of the large-scale trough resulted in cold air damming, which in conjunction with the persistence of this cold reservoir led to continual shallow cold air advection throughout the U.S. Plains. A time series of temperature evolution over the southern Great Plains (Fig. 5.5) shows the gradual decrease in temperatures leading up to 14 February 2021 due to this cold air damming. A cross section through the TPV and southern Great Plains valid at 1200 UTC 12 February 2021 (Fig. 5.6) shows an expansive cold pool maximized beneath the TPV core, with a shallow cold air mass associated with this cold pool extending toward the Gulf of Mexico coast.

The second round of the CAO consisted of larger-scale and deeper tropospheric cold air advection. Following the completion of the TPV merger, the single resultant TPV lingered over the US–Canada border for another day before becoming sheared as it interacted with the polar jet (Fig. 5.4e–f). Around the same time, differential cyclonic vorticity advection and a broad upper-level trough over the Southwest U.S. with differential cyclonic vorticity advection and favorable jet streak dynamics aiding in forcing cyclogenesis and widespread winter precipitation across the southern Great Plains. The increased pressure gradient with deeper vertical depth allowed for more widespread cold air advection from the TPV cold pool across the region, with the CAO peaking in magnitude on 15–16 February 2021 over the southern Great Plains (Fig. 5.5).

5.1.3 Air parcel trajectory analysis

Despite the TPV being situated well to the north of the southern Great Plains, an analysis of backward air parcel trajectories shows a clear link between the TPVs and the peak of the CAO in the southern Great Plains. Backward air parcel trajectories ending in Oklahoma City on 1200 UTC 15 February 2021 (Fig. 5.7a) and 1200 UTC 16 February 2021 (Fig. 5.7b), during the peak of the cold air outbreak, both show the air parcels were initially constrained within the circulation of the TPVs, near the tropopause level, before being ejected equatorward following the TPV merger and descending in altitude as they entered the southern Great Plains.

Given that there were two TPVs with separate origins, a question of interest is whether parcels ending up over the southern Great Plains were equally distributed between both TPVs, or whether they were more strongly associated with one TPV than the other. To quantify the relationship of these parcel trajectories with both TPVs, the parcel coordinates are re-projected into TPV-relative coordinates for both TPVs. Figure 5.8 shows backward parcel trajectories ending in Oklahoma City on 1200 UTC 15 February 2021 in the top row, and 1200 UTC 16 February 2021 in the bottom row, both at 500 m altitude above ground level (AGL). Parcel coordinates relative to TPV B1 are shown on the left, and TPV A on the right. Colors denote time relative to the end of the parcel trajectory, and line widths denote the fraction of the pressure of the parcel between the surface pressure and 2 PVU pressure at the center of the TPV, such that thicker lines denote a parcel altitude near the tropopause level at the vortex core, and thinner lines denote a parcel altitude near the surface.

For both valid times, the parcel trajectory is most strongly linked to TPV B1, exhibiting the closest proximity to that TPV before exiting the circulation of both TPVs and descending in altitude as it moves towards Oklahoma City (Fig. 5.8a,c). This distinction is most noticeable for the parcel trajectory ending 1200 UTC 16 February 2021, as the parcel trajectory is closest to TPV B1 while executing about three cyclonic loops around that TPV (Fig. 5.8c), while only executing about two cyclonic loops around TPV A while being

displaced much further from that TPV (Fig. 5.8d). The parcel appears to exhibit a cyclonic loop around TPV A as an indirect result of being linked more directly to TPV B1, which itself is looping cyclonically around TPV A.

Given that the parcel analysis in Fig. 5.8 is from a deterministic perspective, sensitivity to the exact ending location of the parcel can be examined further from an ensemble of parcel trajectories. HYSPLIT produces a 27-member ensemble of parcel trajectories by applying horizontal offsets within one grid point, and vertical offsets within 0.01 sigma units, to the initial coordinate of the parcel over Oklahoma City. This 27-member ensemble is calculated for parcel trajectories ending at 250 m and 500 m AGL, and transformed into TPV-relative coordinates for both TPVs, ending at 1200 UTC 15 February 2021 (Fig. 5.9) and 1200 UTC 16 February 2021 (Fig. 5.10). Spatially, both analyses show a similar evolution to Fig. 5.8, where parcels are more strongly linked to TPV B1 than TPV A, although a few members can be identified subjectively that exhibit a closer proximity to TPV A than TPV B1.

The results of Figs. 5.9 and 5.10 can be more succinctly analyzed by comparing the percent of parcel trajectory ensemble members whose parcel trajectories are closest to TPV B1 or TPV A and are within 1000 km of that TPV, which is approximately twice the radius of both TPVs. This calculation is performed for parcel trajectory ensembles ending at 1200 UTC 15 February 2021 (Fig. 5.11) and 1200 UTC 16 February 2021 (Fig. 5.12) with ending altitudes of 250 m, 500 m, and 750 m AGL, and is computed through 1200 UTC 12 February 2021, when TPV B1 becomes increasingly asymmetric and the distance between TPVs A and B1 substantially decreases. For trajectories ending at 1200 UTC 15 February 2021 (Fig. 5.11), the percent of members linked to TPV B1 increases leading up to the onset of the TPV merger for parcels ending at 250 m AGL, peaking just over 80%, while only 20% of parcels are linked more closely to TPV A than TPV B. The contrast between both TPVs is even more pronounced for parcel trajectories ending at 1200 UTC 16 February 2021 (Fig. 5.12), as the percent of parcels linked to TPV B1 increases with

time leading up to the TPV merger for all altitudes, while the percent of parcels linked to TPV A decreases with time leading up to the TPV merger. Just before the merger, 100% of parcels ending at 250 m AGL are linked to TPV B, which decreases to about 68% of parcels ending at 750 m AGL.

5.2 MPAS simulation experiments

This section reviews the MPAS simulations performed for this study. The MPAS model configuration is discussed in section 2.3, the TPV modification methodology is described in section 2.3.1, and the methodology specific to this study is discussed in section 2.3.4.

5.2.1 Synoptic overview of simulations

The control simulation was initialized at 0000 UTC 8 February 2021, and was run through 0000 UTC 16 February 2021 to capture the large-scale synoptic and mesoscale TPV evolution and merger of TPVs A and B1 leading up to the peak of the CAO in the southern Great Plains. The synoptic evolution over North America in the control simulation closely follows ERA5 reanalysis through 10 February 2021 (Fig. 5.13a,b), as TPV A stalls near the border between Ontario and Manitoba and TPV B splits into two TPVs, TPV B1 which loops west and continues rotating about TPV A, and TPV B2 which departs eastward towards Atlantic Canada.

The subsequent evolution of the TPV merger somewhat differs from ERA5 reanalysis, as the merger occurs faster in the control simulation than in ERA5 (Fig. 5.13c). The primary difference is with TPV A, which moves westward faster in the control simulation than in ERA5 between 0000 UTC 10 February 2021 and 0000 UTC 11 February 2021, while TPV B1 evolves similarly to ERA5, leading to a greater proximity of both TPVs in both the control simulation than in ERA5. As the TPV merger continues through 14 February 2021 (Fig. 5.13d,e), the earlier merger of the TPVs is followed by a quicker shearing of the resultant vortex, and a less amplified downstream ridge over Quebec, in

the control simulation than in ERA5. By 0000 UTC 15 February 2021 (Fig. 5.13f), the TPV merger is complete with the resultant vortex becoming sheared as it interacts with the midlatitude jet in both the control simulation and ERA5. The primary difference at this valid time is a weaker vortex in the control simulation than in ERA5.

Differences between the control simulation and ERA5 are mostly constrained to the TPV merger and its subsequent evolution post-merger. The large-scale synoptic evolution over North America, and the 850-hPa temperature pattern, evolves similarly to ERA5, lending confidence to the control simulation being a sufficient basis for the TPV intensity modification experiments by both offering sufficient lead time to modify both TPVs and broadly capturing the observed large-scale synoptic evolution of the CAO.

Six simulations with modified TPV intensities were performed, with the TPV modification algorithm applied over a 24-h period between 1200 UTC 8 February 2021 and 1200 UTC 9 February 2021. Three simulations intensified the TPV by applying potential temperature tendencies of -15 K day^{-1} (-15K), -10 K day^{-1} (-10K), and -5 K day^{-1} (-5K), and three simulations weakened the TPV by applying potential temperature tendencies of $+5 \text{ K day}^{-1}$ (+5K), $+10 \text{ K day}^{-1}$ (+10K), and $+15 \text{ K day}^{-1}$ (+15K). At the end of the TPV modification window valid 1200 UTC 9 February 2021 (Fig. 5.14), the most pronounced impact is noticeable in the -15K, -10K and -5K simulations, as both TPV A and B1 are substantially stronger than the control simulation. The +5K, +10K and +15K simulations have weaker TPVs than the control simulation, but due to the constraint on the TPV outer edge imposed to prevent the vortex from completely dissipating when applying a heating tendency, the differences in the intensities of both TPVs in these three simulations relative to the control simulation are smaller than those of the -5K, -10K and -15K simulations. A subtle difference that cannot be easily identified in Fig. 5.14, but becomes important in later forecast hours, is that simulations with stronger TPVs have a slightly more zonal orientation of the northern edge of TPV B1 than simulations with weaker TPVs.

More substantial differences emerge in the rate of the TPV merger between the simulations by forecast hour 84, valid 1200 UTC 11 February 2021 (Fig. 5.15). Simulations with stronger TPVs have a faster TPV merger than simulations with weaker TPVs, with differences between simulations most noticeable in the -10K and -15K simulations where TPV B1 retains a relatively axisymmetric structure as TPV A rapidly becomes sheared while merging into TPV B1. This evolution is not observed in simulations with weaker TPVs, nor is this observed in ERA5 analyses. Another notable difference between all simulations is the location and intensity of a PV lobe extending northwest of both TPVs into British Columbia, with its formation associated with some deformation of both TPVs. Its location and intensity are similar between the control, +5K, +10K, and +15K simulations, while the -5K, -10K, and -15K simulations sequentially have a more intense PV lobe located farther southeast, indicating less of a spatial displacement from TPV B1. A subjective evaluation of the 2 PVU potential temperature field between all members indicates this is related to variability in the zonal orientation of the northern gradient of TPV B1 discussed in Fig. 5.14.

By forecast hour 144, valid 0000 UTC 14 February 2021 (Fig. 5.16), the TPV merger is completed across all simulations. The -15K simulation has the strongest TPV, and the +15K simulation has the weakest TPV. Differences in the PV lobe between simulations discussed in Fig. 5.15 exert a larger impact on the large-scale synoptic evolution by this time. Variability between the +15K, +10K, and +5K simulations, which are all consistent in the evolution of this PV lobe, is mostly restricted to the rate of merger of the TPVs and associated slight differences in the downstream low-amplitude ridge over Quebec. In contrast, among the control, -5K, -10K, and -15K simulations, a more intense PV lobe is associated with a more meridional waveguide east of the Rockies, a more meridional orientation of the merged TPV, and a more amplified cyclone beginning to form over the eastern United States. In terms of sensible impacts, among the control, -5K, -10K, and -15K simulations, a stronger and more meridionally oriented merged TPV is associated with

the core of the cold pool associated with the TPV displaced farther poleward into Canada at this valid time. Accordingly, simulations with a stronger TPV are warmer over the southern Great Plains, despite being associated with a stronger cold pool, than simulations with a weaker TPV.

Spatial differences between all simulations can be succinctly visualized through the standard deviation of 500-hPa geopotential height of all simulations at select valid times (Fig. 5.17). As expected, the largest variability at the end of the TPV modification window valid 1200 UTC 9 February 2021 (Fig. 5.17a) is within the core of the TPVs, due to the upper-tropospheric heating and cooling tendencies applied to modify the intensity of the TPVs. By 11 February 2021 (Fig. 5.17b), however, the greatest 500-hPa geopotential height spread is in the region between TPVs A and B1, denoting variability in their merger rate, and in the amplitude of the PV lobe extending northwest of both TPVs. Over the subsequent two days (Fig. 5.17c,d), variability in the PV lobe evolution leads to timing and amplitude variability of a shortwave trough entering the western United States, and variability in the orientation and location of the geopotential height gradient south of both TPVs, which translates at later valid times into variability in the amplitude of a broad ridge over eastern North America (Fig. 5.17e,f). The largest 500-hPa geopotential height variability at later forecast lead times is directly upstream of the TPV post-merger (Fig. 5.17e,f), as simulations with stronger TPVs and a faster TPV merger have the resultant merged vortex situated northwest of simulations with weaker TPVs and a slower TPV merger.

From a dynamic tropopause perspective, the end of the TPV modification window valid 1200 UTC 9 February 2021 (Fig. 5.18a) exhibits a similar but more pronounced pattern than the 500-hPa geopotential height standard deviation (Fig. 5.17a), with the greatest magnitude over the centers of both TPVs as would be expected after modifying the intensities of both TPVs. While subsequent forecast hours maintain some level of 2 PVU potential

temperature spread within the cores of both TPVs, the magnitudes of 2 PVU potential temperature spread at the edges of the vortices grows in magnitude by 11 February 2021 (Fig. 5.18b), and eventually overtakes the cores of both TPVs as the largest regions of variability in the ensemble (Fig. 5.18c).

5.2.2 southern Great Plains temperature analysis

2-meter temperature during the first 24 hours of the simulations evolves nearly identically to ERA5 (Fig. 5.19). Subsequently, all simulations develop a systematic warm bias relative to ERA5, reaching a magnitude of 5°C as soon as 60-hour forecast lead time. This warm bias persists through the remainder of all simulations, despite these simulations correctly capturing the observed decrease in 2-meter temperature and the timing of the peak of the CAO. Differences in 2-meter temperature over the southern Great Plains between simulations begin to become distinguishable on 11 February 2021, around forecast hours 84–96, as the differences in the evolution of the PV lobe discussed in the previous section begin to influence the downstream waveguide over the United States. By 12 February 2021, stronger TPVs are associated with warmer temperatures in the southern Great Plains, while weaker TPVs are associated with colder temperatures in the southern Great Plains.

The warm bias shown in Fig. 5.19 resembles a surface warm bias in operational NWP model forecasts during the CAO, although these biases emerged in shorter forecast lead times in operational models relative to the MPAS simulations performed in this experiment. A comparison of modeled vertical profiles from the operational High-Resolution Rapid Refresh (HRRR) (Dowell et al. 2022) model (Fig. 5.20a) and the MPAS control simulation (Fig. 5.20b) against radiosonde observations shows that at 36-h forecast lead time, the HRRR exhibited a low-level warm bias while the MPAS was much closer to observations, while still exhibiting a slight warm bias. By forecast hour 84 (Fig. 5.21), however, a similar magnitude low-level warm bias is evident in the MPAS control simulation. Spatially, a

warm 2-meter temperature bias is evident in the short range of the MPAS control simulation across the Great Plains, with the greatest warm bias in the lee of the Rockies (Fig. 5.22).

Failure of models to adequately resolve the magnitude of shallow inversions, and in turn exhibit a warm low-level bias during cold air damming regimes, has been documented in cases of frozen precipitation associated with cold air damming (e.g., Ikeda et al. 2017). More recently, data from the Winter Precipitation Type Research Multiscale Experiment (WINTRE-MIX) field campaign (Minder et al. 2023) found a similar low-level warm bias in a cold air damming environment as a result of models failing to capture the magnitude of the shallow inversion, while modeled temperatures in the warm layer aloft were much closer to sounding data recorded during the field campaign.

5.3 Discussion

The primary result of the modified TPV intensity simulations for the February 2021 CAO is that the relationship between TPV intensity and temperatures in the southern Great Plains is non-linear. Figure 5.25 shows an idealized schematic of the relationship between TPV intensity and CAO evolution, highlighting the competing influence between a stronger cold pool magnitude with a stronger TPV magnitude, and the control, -5K, -10K, and -15K simulations exhibiting a greater poleward displacement of the cold pool as a stronger PV lobe over western Canada is associated with a more meridional orientation of the merged TPV and the overall North American waveguide. The end result is that the control simulation is the coldest of all seven simulations in the southern Great Plains.

Notably, this finding differs from the results of Lillo et al. (2021), who performed a similar TPV intensity modification study but on a single fast-moving TPV. In their study, a stronger TPV exiting Canada towards the Great Lakes was associated with colder temperatures across large swaths of the eastern United States, while a weaker TPV was associated

with warmer temperatures across the same regions in the United States. This linear relationship between TPV intensity and CAO magnitude was not found in this study. It is speculated that the presence of multiple TPVs, their interaction and merger, and their associated PV lobe in western Canada add more degrees of freedom that result in a more complex synoptic-scale evolution of the CAO than a single, fast-moving TPV where a more intense TPV and associated cold pool would quickly and directly translate into stronger cold air advection downstream of the TPV. Accordingly, this result fails to support hypothesis C2, as stronger TPVs do not lead to a stronger cold outbreak in the southern Great Plains than weaker TPVs.

A unique aspect of this CAO was the merger of both TPVs. Hypothesis C1, which expected stronger TPVs to merge at a faster rate than weaker TPVs, is supported by these MPAS simulations (Recall Fig. 5.15). This experiment, however, does not offer clear evidence in support of hypothesis C3, which stated a faster TPV merger would result in the merged TPV propagating downstream earlier, and resulting in a shorter duration of the CAO in the southern Great Plains, relative to a slower TPV merger. Simulations with weaker TPVs and accordingly a slower TPV merger do end up with a more zonally oriented merged TPV orientation than simulations with stronger TPVs (Recall Fig. 5.16), but as discussed in section 5.2.1, this difference appears to be more related to synoptic-scale differences in the waveguide originating from variability in the TPV lobe over western Canada.

From an operational forecasting perspective, the results of this experiment can be compared to the 51-member ECMWF Ensemble Prediction System (EPS), which is produced in real time with perturbations applied to both ICs and parameterization tendencies. Applying ESA to the EPS allows to assess sensitivity of the CAO evolution to additional sources of forecast variability that are not solely constrained to variability in the intensity the TPVs, while providing an analysis that can be readily made available to forecasters in real time. A forecast initialization of 0000 UTC 7 February 2021 was selected to allow the ensemble to

develop sufficient spread in the TPV evolution over Canada in the short to medium forecast range, with a valid time of 0000 UTC 15 February 2021 as the CAO nears its peak in the southern Great Plains.

The response function J is constructed by averaging 2-meter air temperatures over the southern Great Plains domain for each ensemble member, calculating the difference of each ensemble member from the ensemble mean, then normalizing by the ensemble standard deviation. Based on sensitivity of the southern Great Plains averaged temperature to 500-hPa geopotential height valid 0600 UTC 11 February 2021 (Fig. 5.23), colder temperatures in the southern Great Plains at 0000 UTC 15 February 2021 are associated with a stronger and more poleward shifted PV lobe over British Columbia, higher heights to the south of the merging TPVs over the Dakotas, a less amplified ridge downstream of the shortwave trough approaching western California, and higher heights upstream of a trough axis over the southwest United States, valid at 0600 UTC 11 February 2021. Repeating the ESA computation with sensitivity to 2 PVU potential temperature at 0600 UTC 11 February 2021 (Fig. 5.24), the largest sensitivity is associated with the evolution of the PV lobe over northern Manitoba and British Columbia, which is linked to the northern gradient of TPV B1.

These results align with ESA analyses applied to aircraft reconnaissance missions studying forecast sensitivity of atmospheric rivers (ARs), as while these are different phenomena from TPVs, a similar conclusion is that the largest source of forecast sensitivity is not from areas where the magnitude of the feature of interest is largest, which is 2 PVU potential temperature in this case and fields such as humidity, wind and moisture transport in AR cases, but rather where horizontal gradients in these fields are large (e.g., Reynolds et al. 2019; Zheng et al. 2024). In this case, the gradient to the north of TPV B1 is a large source of forecast uncertainty and sensitivity, and variability in this gradient can be associated with variability in TPV asymmetry and intensity, implying that TPV intensity variability can be an indirect source of forecast sensitivity.

An overview of deterministic and ensemble backward air parcel trajectories from Oklahoma City on 1200 UTC 15 February 2021 and 1200 UTC 16 February 2021 reveal that most of these parcels are associated with TPV B1, being constrained within its circulation leading up to the TPV merger before being ejected equatorward and descending in altitude. When combined with the synoptic overview of the ensemble members and the ESA analysis of the EPS ensemble, this indicates that TPV B1 had a greater direct role on the evolution of the CAO in the southern Great Plains than TPV A. However, this does not mean that TPV A had little or no role in the evolution of the CAO, as the location and intensity of TPV A does affect the location and merger rate of TPV B1. Future work more closely isolating the role of TPV B1 and TPV A would be needed to more conclusively determine the exact contribution of each TPV to the CAO evolution in the southern Great Plains.

5.3.1 Overview of additional simulations

The selection of the TPV modification domain shown in Fig. 2.9 allowed to modify TPV B1 without modifying both the PV filament over Alberta and the PV filament that split off into TPV B2. A set of ensemble simulations was also ran for this experiment that did not include this distinction, while also using a slightly different TPV modification window between 0000 UTC 8 February 2021 and 0000 UTC 9 February 2021. The TPVs in this ensemble evolved similarly to the main experiment through 11 February 2021, though a major difference was the evolution of TPV B2, which exhibited an inflection point between simulations with weaker TPVs, in which case TPV B2 looped back to the west and became sheared off as it interacted with TPV A, and simulations with stronger TPVs, in which case TPV B2 rapidly departed to the east into Atlantic Canada. The impact of this inflection point grew into later forecast hours such that differences between simulations in the CAO evolution over the southern Great Plains became nonlinear. The temporal evolution of temperatures averaged over the southern Great Plains (Fig. 5.26) also shows a larger and

earlier onset of deviation among simulations from the control simulation relative to the main experiment (Fig. 5.19).

Two additional factors that have not been discussed yet are the importance of choice of ICs and model domain selection. Earlier experiments, not shown in this chapter, used a global MPAS mesh with a global 60-km mesh and a gradual refinement region towards a higher resolution 15-km mesh centered over North America. The control simulation in this experiment deviated from ERA5 reanalysis more rapidly than the limited-area MPAS simulation. On the synoptic scale, there are two primary differences between the regional and global MPAS control simulations. The global simulation had smaller separation between both TPVs and was faster to merge them compared to the regional simulation, though both simulations were too far west with the positioning of TPV A. Following the completion of the TPV merger, the global simulation had a more meridional orientation to the resultant vortex, in addition to more amplified ridging upstream of the vortex, than the regional simulation. This difference resulted in the TPV expanding farther south into the Great Lakes and Ohio Valley in the global simulation than the regional simulation. In terms of sensible impacts, the end result was the peak of the CAO over the southern Great Plains in all global simulations was about a day earlier than both regional simulations and observations.

The second important source of sensitivity is the choice of ICs. This experiment used ERA5 reanalysis ICs, as given the selected forecast initialization time, this resulted in a realistic synoptic evolution. Earlier experiments, not shown here, used the GFS for ICs. What appeared to be a potentially spurious minimum in 2 PVU potential temperature within the core of the precursor to TPV B in the GFS ICs propagated downstream with the simulation and grew in magnitude, eventually resulting in TPV B erroneously becoming stronger than TPV A. Towards the peak of the CAO, the evolution of the CAO and overall North American waveguide became substantially different from both the ERA5 IC simulation and ERA5 reanalysis.

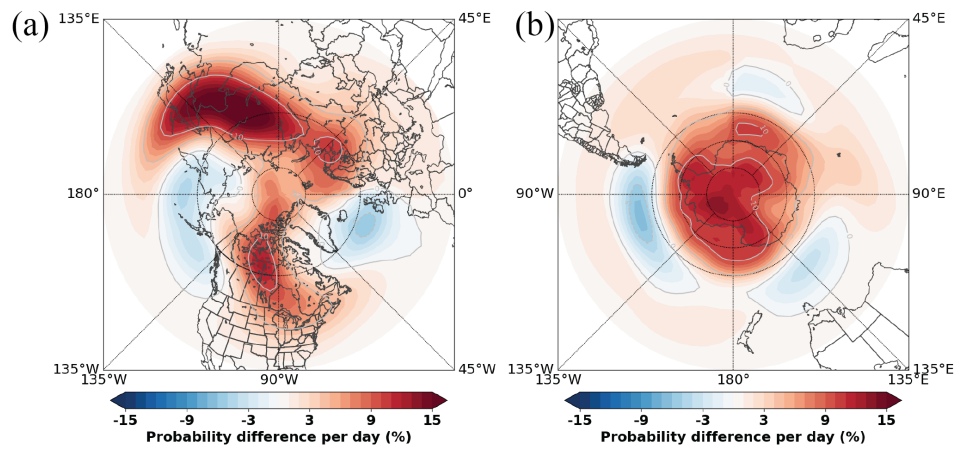


Figure 5.1: Difference in TPV probability per day, defined as the average probability of at least one TPV being located within 555 km of the corresponding point per day over the respective period (i.e., DJF or JJA), for winter minus summer months in the (a) Northern Hemisphere, where DJF denotes winter and JJA denotes summer, and (b) Southern Hemisphere, where JJA denotes winter and DJF denotes summer.

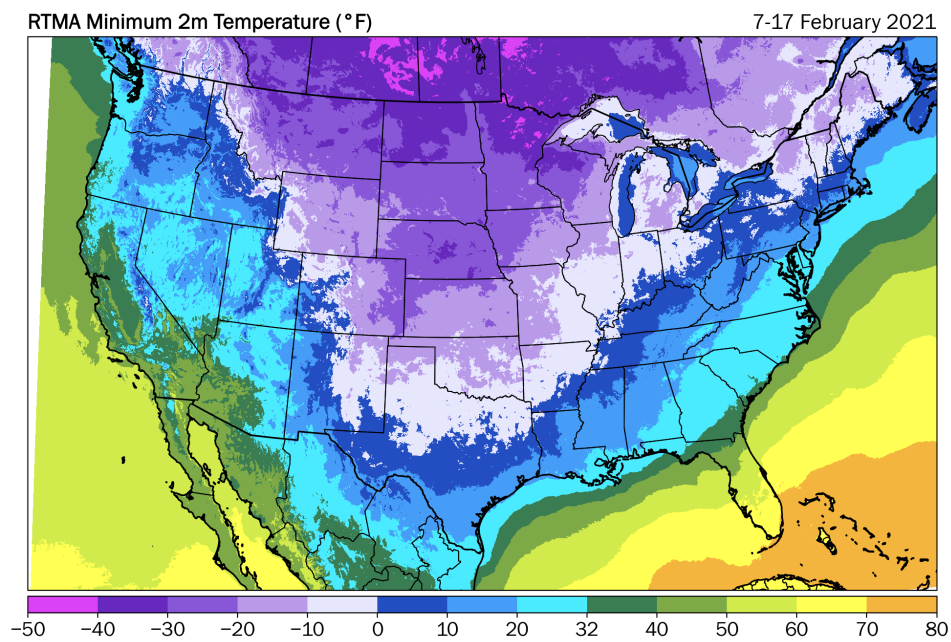


Figure 5.2: Minimum 2-meter temperatures (°F) between 0000 UTC 17 February 2021 and 2300 UTC 17 February 2021, as derived from the hourly 2.5 km horizontal resolution Real Time Mesoscale Analysis (RTMA; De Pondeca et al. 2011).

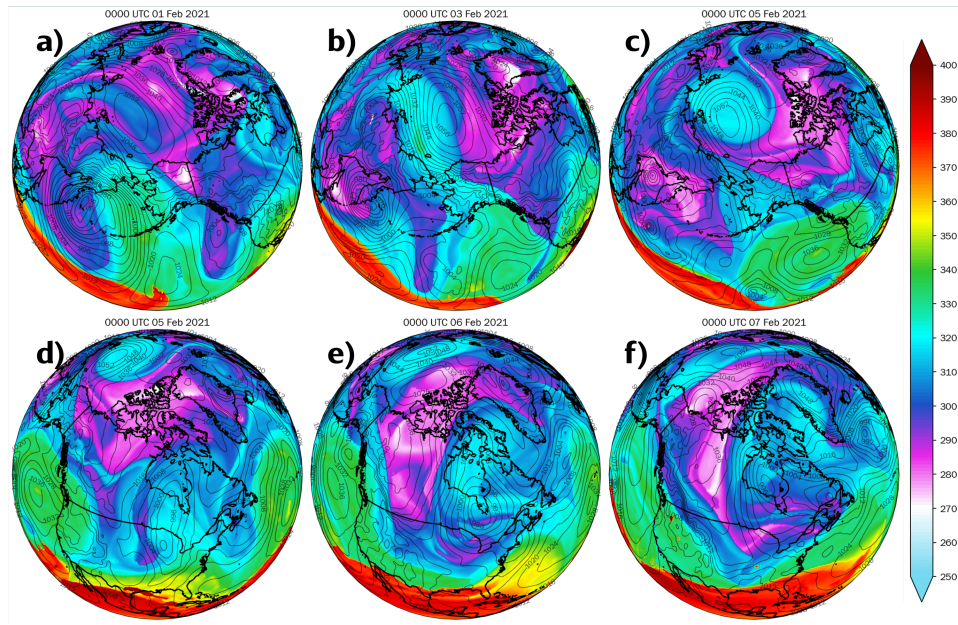


Figure 5.3: ERA5 2 PVU potential temperature (K) in fill and mean sea level pressure (hPa) in contours. Top row shows the evolution of the Arctic anticyclone, valid at (a) 0000 UTC 1 February 2021, (b) 0000 UTC 3 February 2021, and (c) 0000 UTC 5 February 2021. Bottom row shows the evolution of the Greenland anticyclone, valid at (d) 0000 UTC 5 February 2021, (e) 0000 UTC 6 February 2021, and (f) 0000 UTC 7 February 2021.

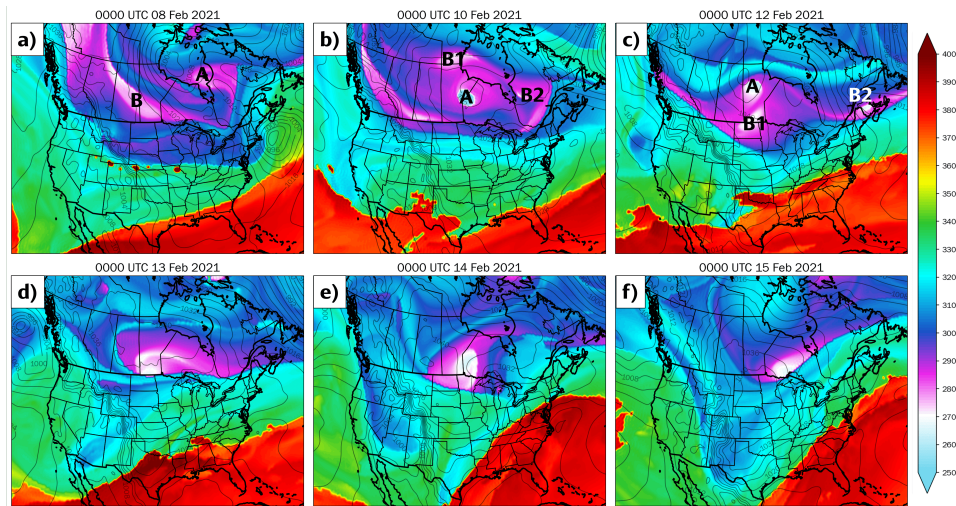


Figure 5.4: ERA5 2 PVU potential temperature (K) in fill and mean sea level pressure (hPa) in contours. Frames are valid at (a) 0000 UTC 8 February 2021, (b) 0000 UTC 10 February 2021, (c) 0000 UTC 12 February 2021, (d) 0000 UTC 13 February 2021, (e) 0000 UTC 14 February 2021, and (f) 0000 UTC 15 February 2021.

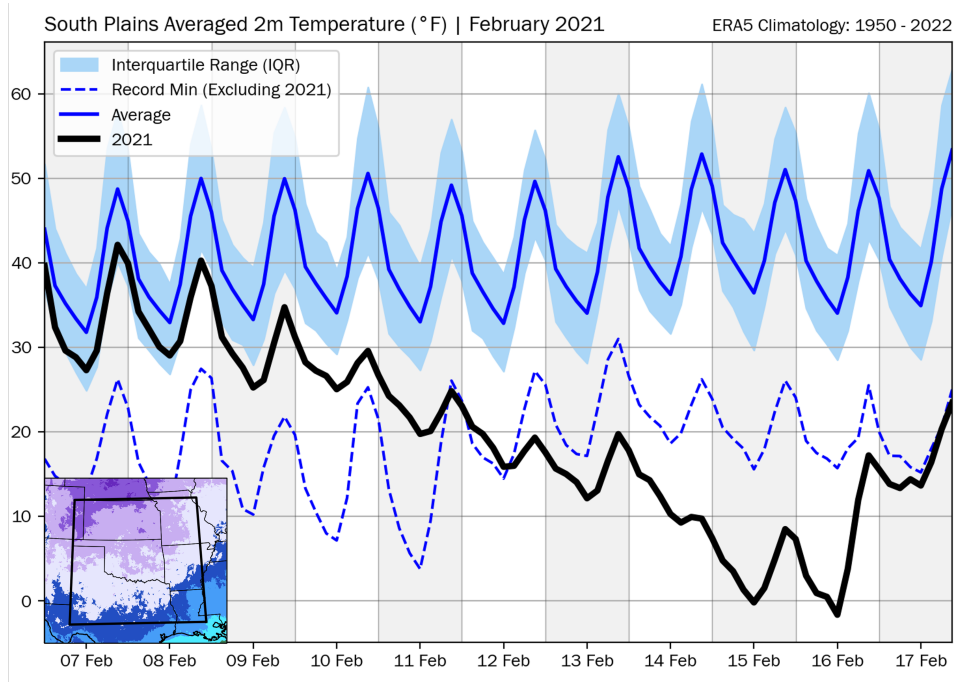


Figure 5.5: Time series of ERA5 2-meter temperatures averaged over the denoted southern Great Plains domain from February 2021 (black line), compared against ERA5 1950–2022 average (blue line), interquartile range (blue shading), and minimum record excluding 2021 (dashed blue line).

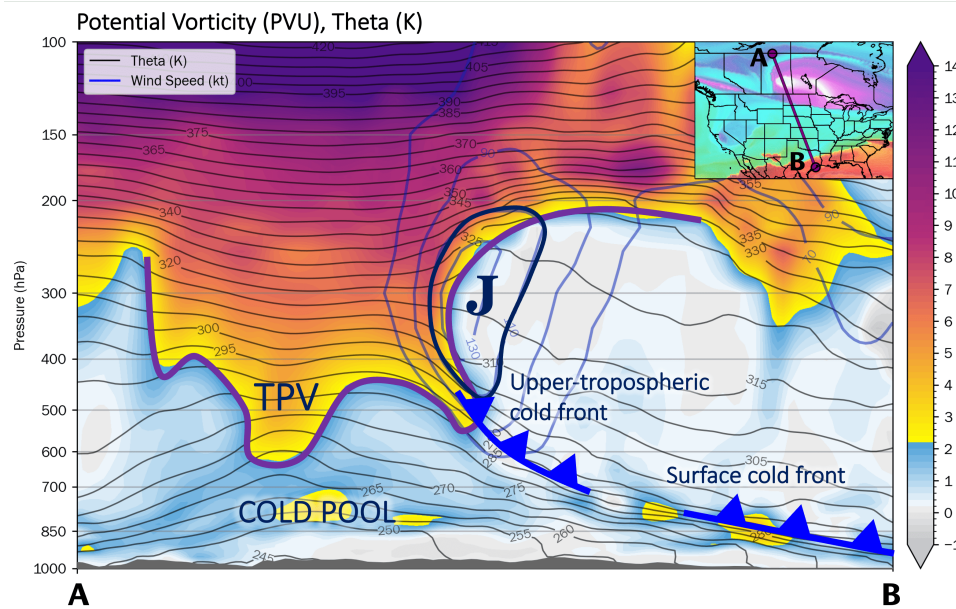


Figure 5.6: ERA5 Cross section of EPV (PVU; fill), potential temperature (K; black contours) and wind speed isotachs (knots; blue contours), valid at 1200 UTC 12 February 2021. Upper right inset shows 2 PVU potential temperature valid at the same time.

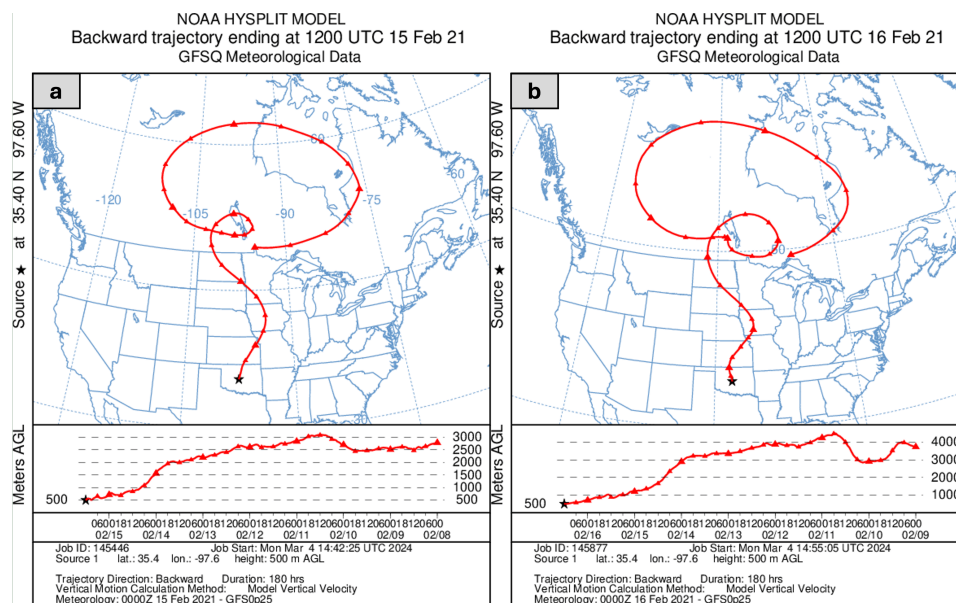


Figure 5.7: Backward air parcel trajectories ending in Oklahoma City at 500 m altitude on (a) 1200 UTC 15 February 2021, and (b) 1200 UTC 16 February 2021. Images generated from the HYSPLIT program (Stein et al. 2015).

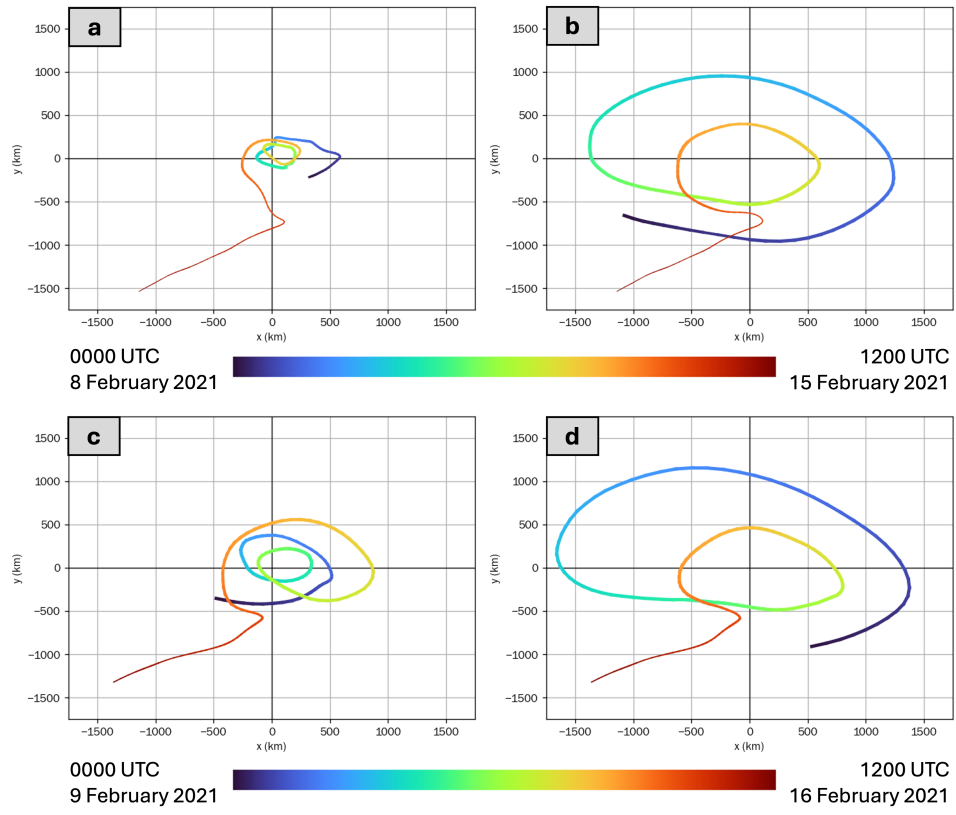


Figure 5.8: Backward parcel trajectories ending in Oklahoma City on (a, b) 1200 UTC 15 February 2021 in the top row, and (c, d) 1200 UTC 16 February 2021 in the bottom row, all ending at 500 m altitude. Parcel coordinates relative to TPV B1 are shown on the left (a, c), and TPV A on the right (b, d). Colors denote time relative to the end of the parcel trajectory, and line widths denote the fraction of the pressure of the parcel between the surface pressure and 2 PVU pressure at the center of the TPV, such that thicker lines denote a parcel altitude near the tropopause level at the vortex core, and thinner lines denote a parcel altitude near the surface.

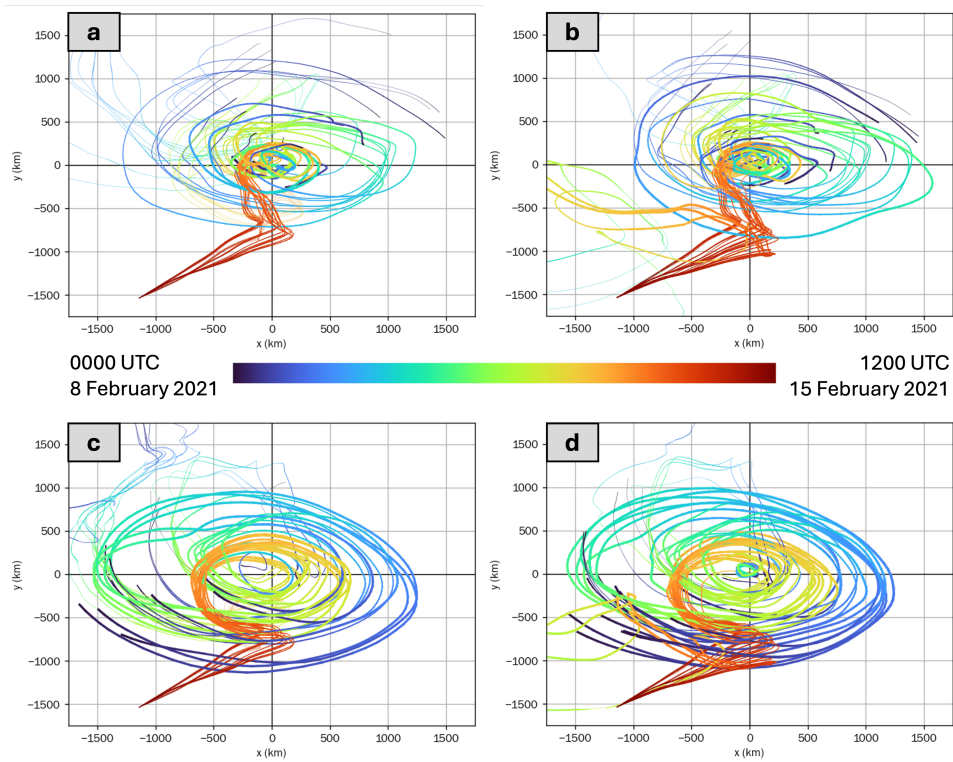


Figure 5.9: Backward air parcel trajectory ensembles ending in Oklahoma City on 1200 UTC 15 February 2021, at 250 m altitude on the left (a, c) and 500 m altitude on the right (b, d). Parcel coordinates relative to TPV B1 are shown on the top (a, b), and TPV A on the bottom (c, d). Colors and line thickness follow the same convention as Fig. 5.8.

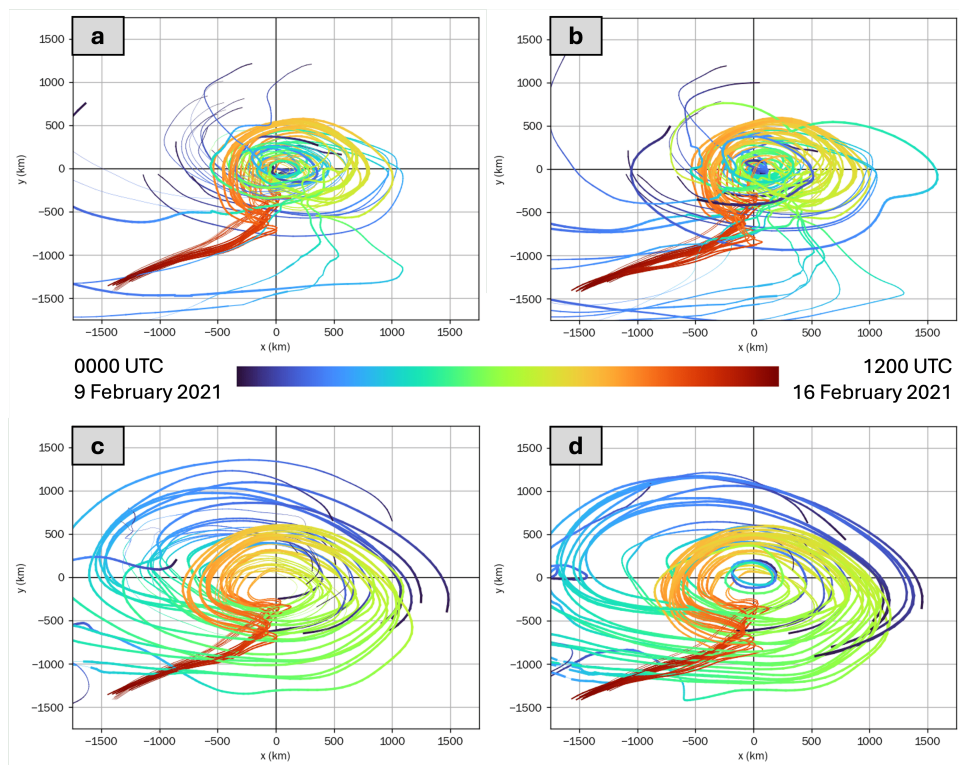


Figure 5.10: Same as Fig. 5.9, but ending at 1200 UTC 16 February 2021.

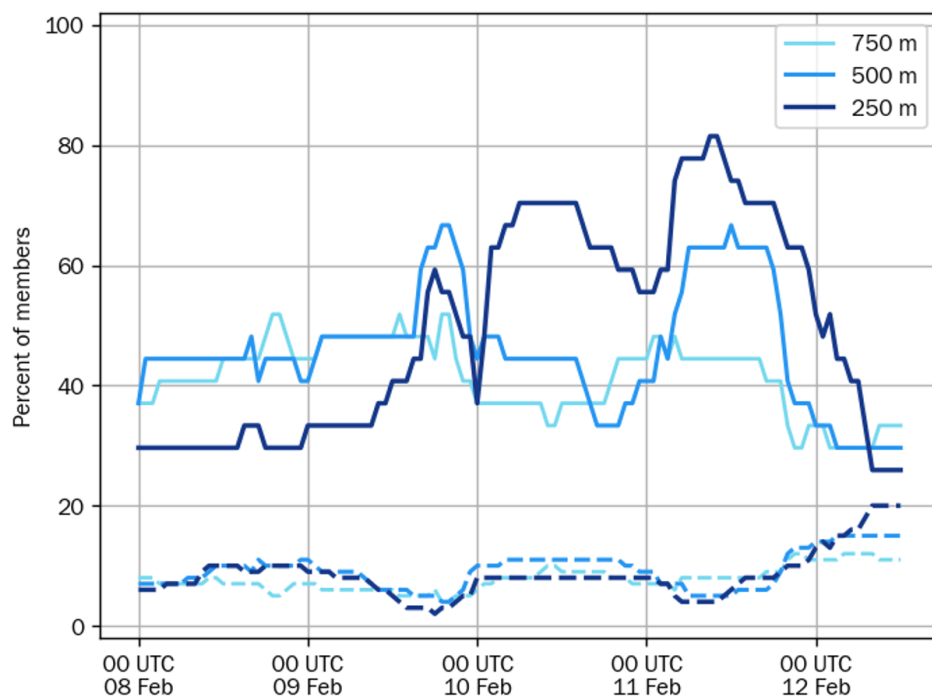


Figure 5.11: Percent of HYSPLIT ensemble members ending at 1200 UTC 15 February 2021 at 250 m altitude above ground level (cyan), 500 m altitude (light blue), and 750 m altitude (dark blue), whose parcel trajectories are closest to TPV B1 (solid lines) or TPV A (dashed lines) and within 1000 km of that TPV.

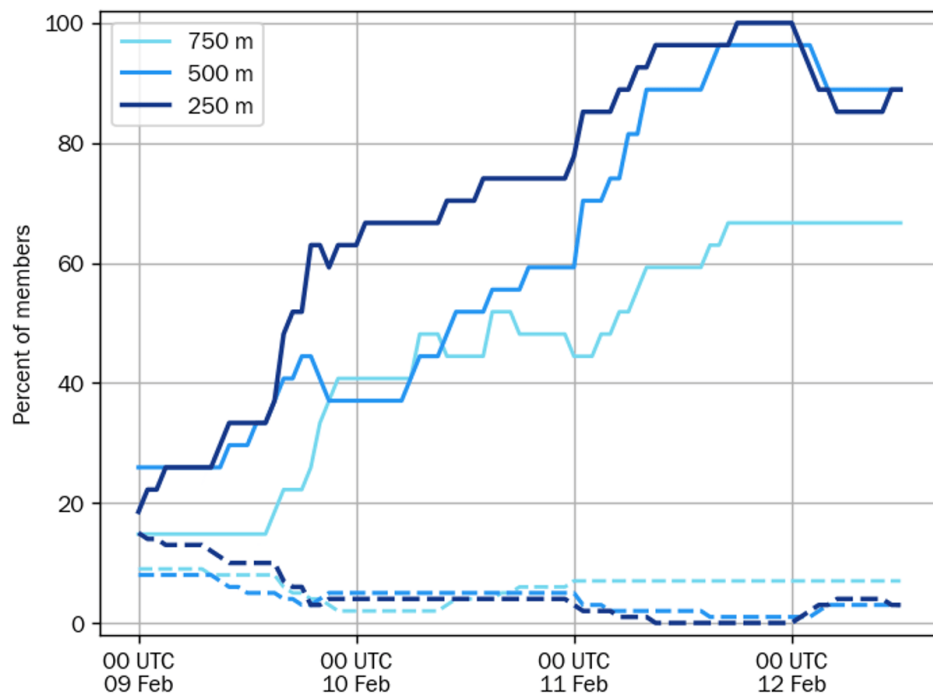


Figure 5.12: Same as Fig. 5.11, but ending at 1200 UTC 16 February 2021.

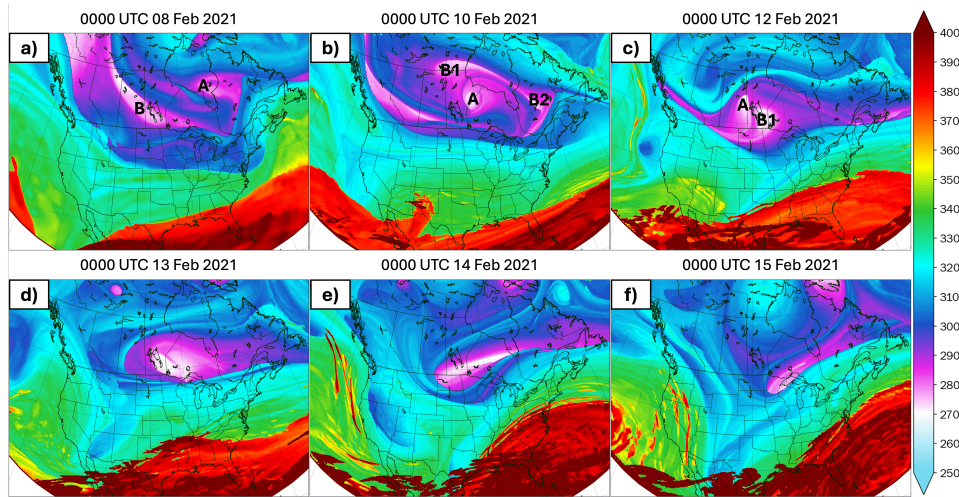


Figure 5.13: Control simulation 2 PVU potential temperature (K), valid at (a) 0000 UTC 8 February 2021, (b) 0000 UTC 10 February 2021, (c) 0000 UTC 12 February 2021, (d) 0000 UTC 13 February 2021, (e) 0000 UTC 14 February 2021, and (f) 0000 UTC 15 February 2021.

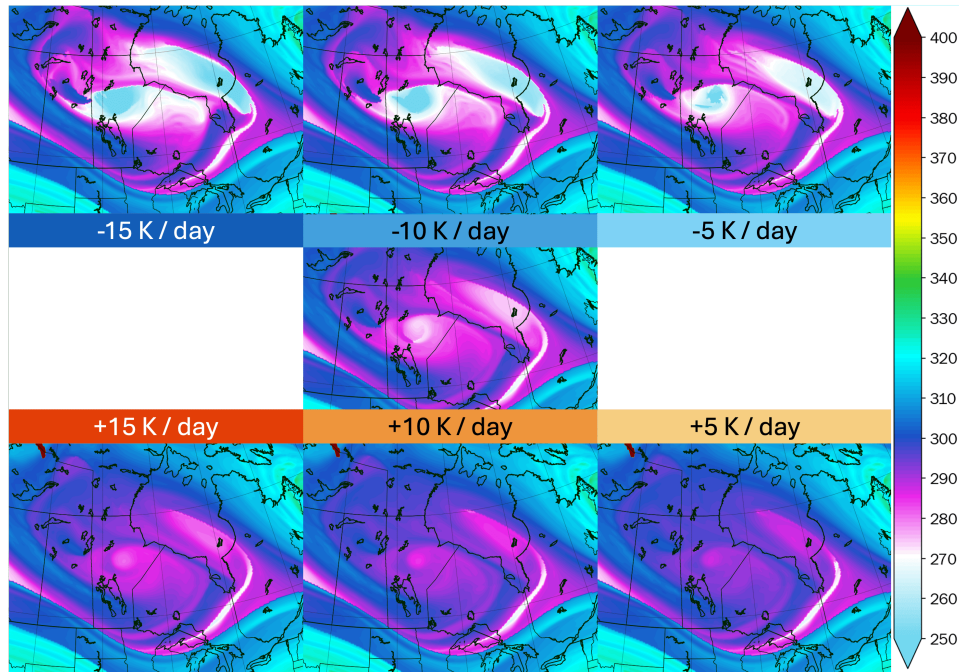


Figure 5.14: 2 PVU potential temperature (K) valid at 1200 UTC 9 February 2021, at the end of the TPV modification period compared across all simulations. The control simulation is shown in the center.

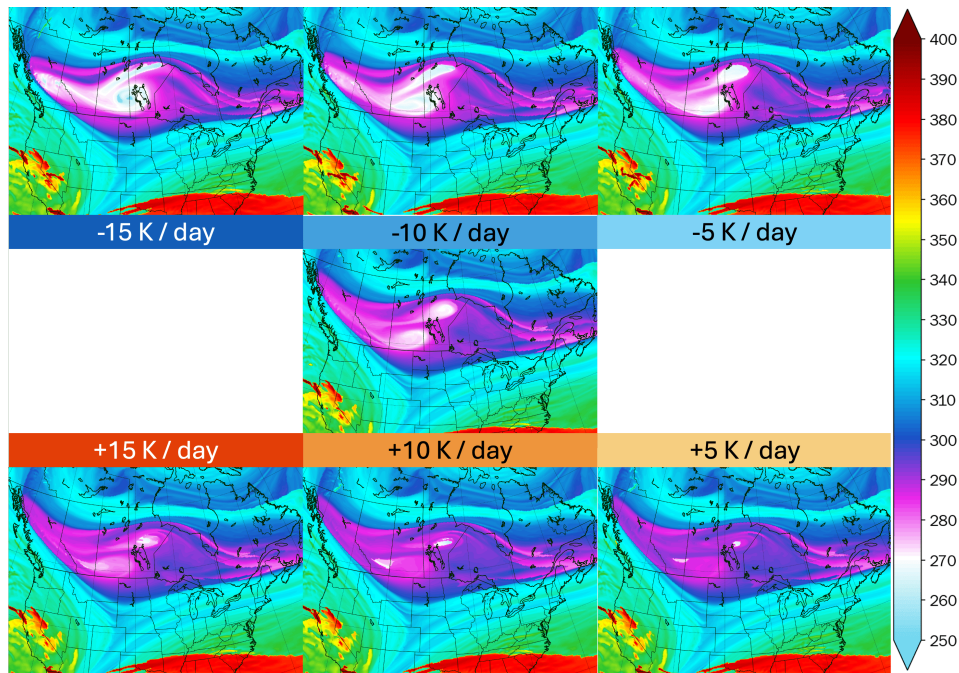


Figure 5.15: Same as Fig. 5.14, but valid at 1200 UTC 11 February 2021. TPV B1 is located to the south, and TPV A to the north, in all panels.

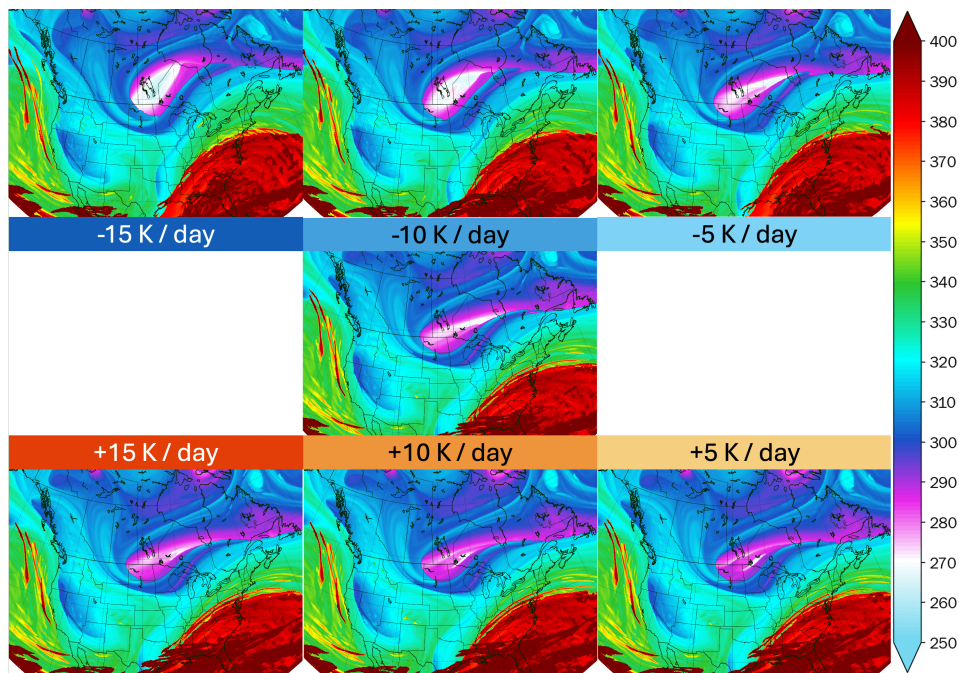


Figure 5.16: Same as Fig. 5.14, but valid at 0000 UTC 14 February 2021.

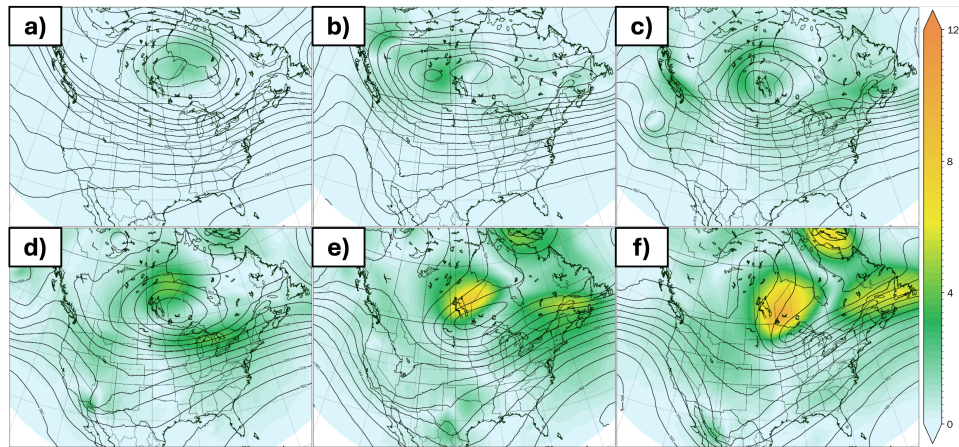


Figure 5.17: MPAS ensemble standard deviation in 500-hPa geopotential height (dam) in fill, and control member 500-hPa geopotential height (dam) in contours, valid (a) 1200 UTC 9 February 2021, (b) 0000 UTC 11 February 2021, (c) 0000 UTC 12 February 2021, (d) 0000 UTC 13 February 2021, (e) 0000 UTC 14 February 2021, and (f) 0000 UTC 15 February 2021.

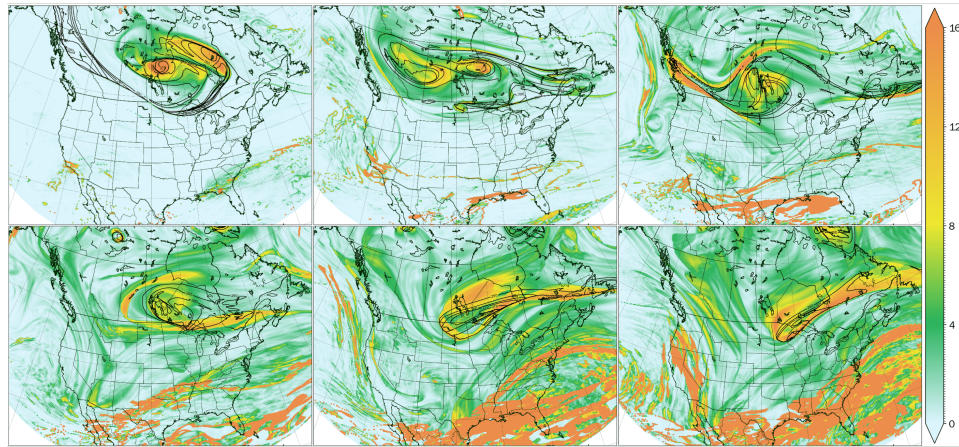


Figure 5.18: MPAS ensemble standard deviation in 2 PVU potential temperature (K) in fill, and control member 2 PVU potential temperature (K) contoured every 5 K up to 290 K, valid (a) 1200 UTC 9 February 2021, (b) 0000 UTC 11 February 2021, (c) 0000 UTC 12 February 2021, (d) 0000 UTC 13 February 2021, (e) 0000 UTC 14 February 2021, and (f) 0000 UTC 15 February 2021.

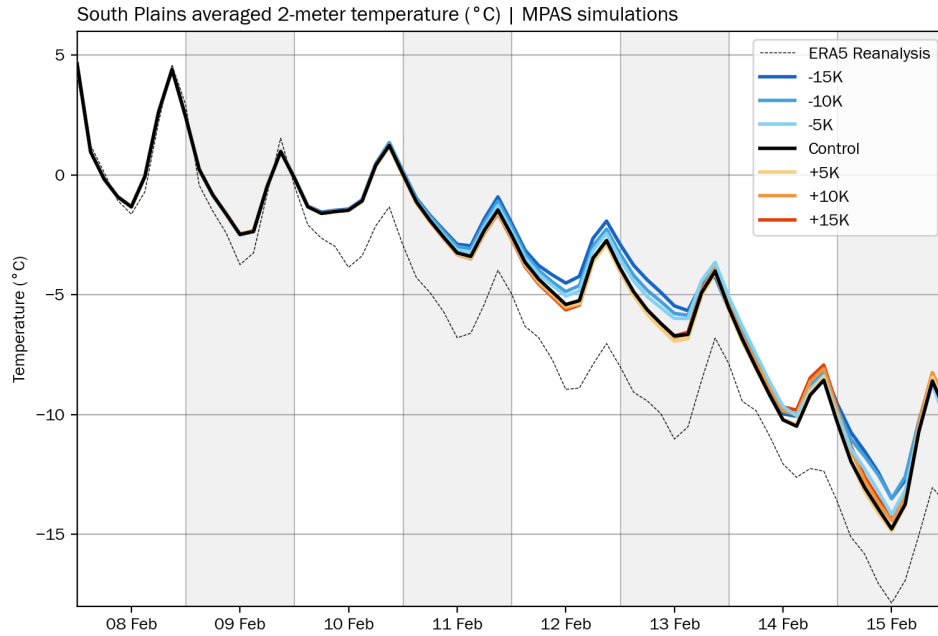


Figure 5.19: 2-meter temperature ($^{\circ}\text{C}$) averaged over the southern Great Plains domain for ERA5 (dotted black line) and the various MPAS simulations performed for this experiment.

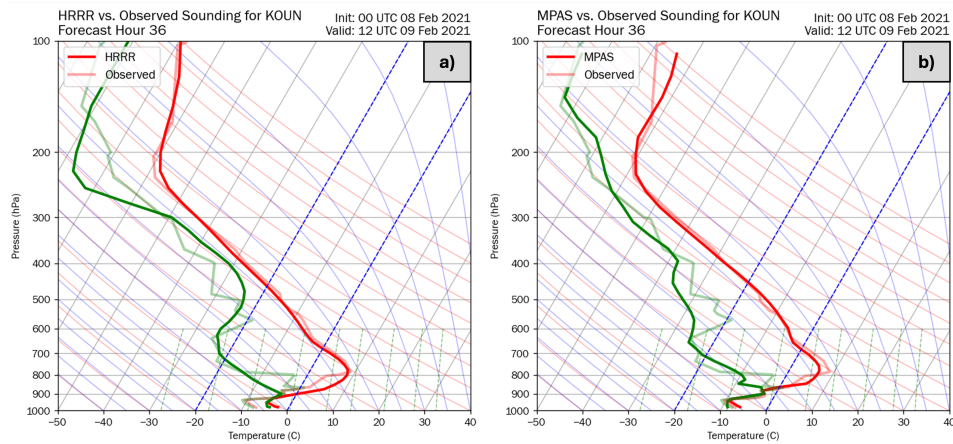


Figure 5.20: Vertical temperature (red) and dewpoint (blue) profiles for (a) HRRR and (b) MPAS control simulation in solid lines, with radiosonde observations in transparent colors. Both simulations were initialized at 0000 UTC 8 February 2021, valid at 1200 UTC 9 February 2021.

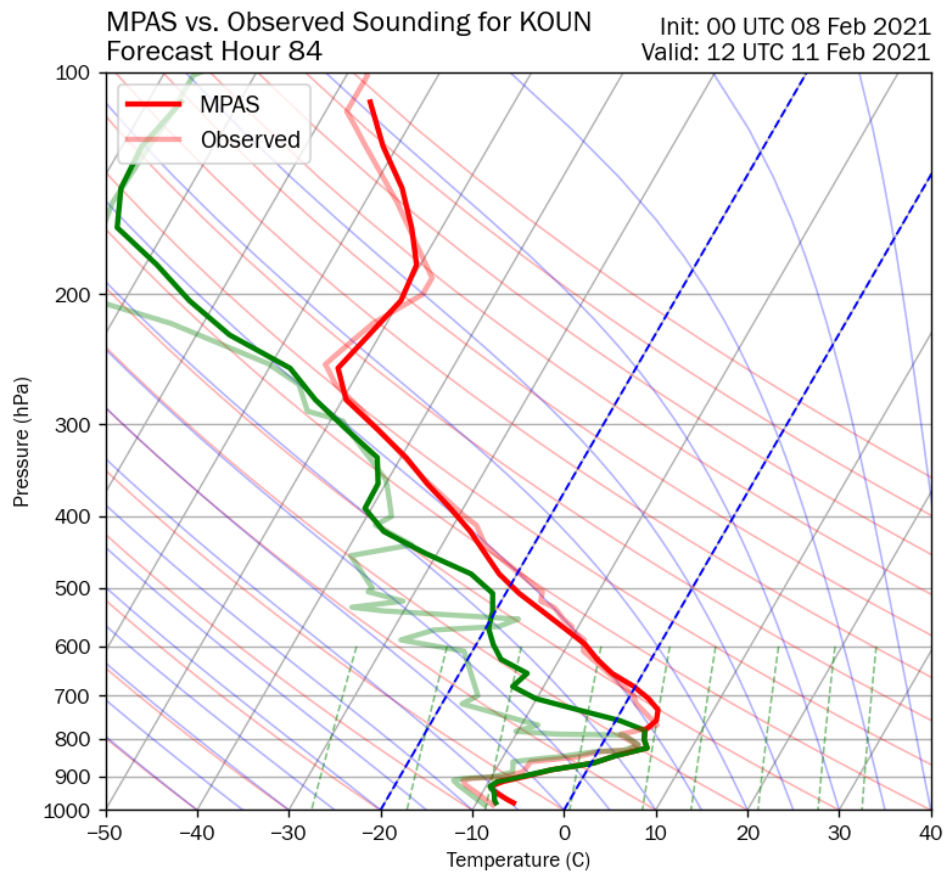


Figure 5.21: Vertical (red) and dewpoint (blue) profiles for the MPAS control simulation in solid lines, with radiosonde observations in transparent color, initialized at 0000 UTC 8 February 2021 and valid at 1200 UTC 11 February 2021.

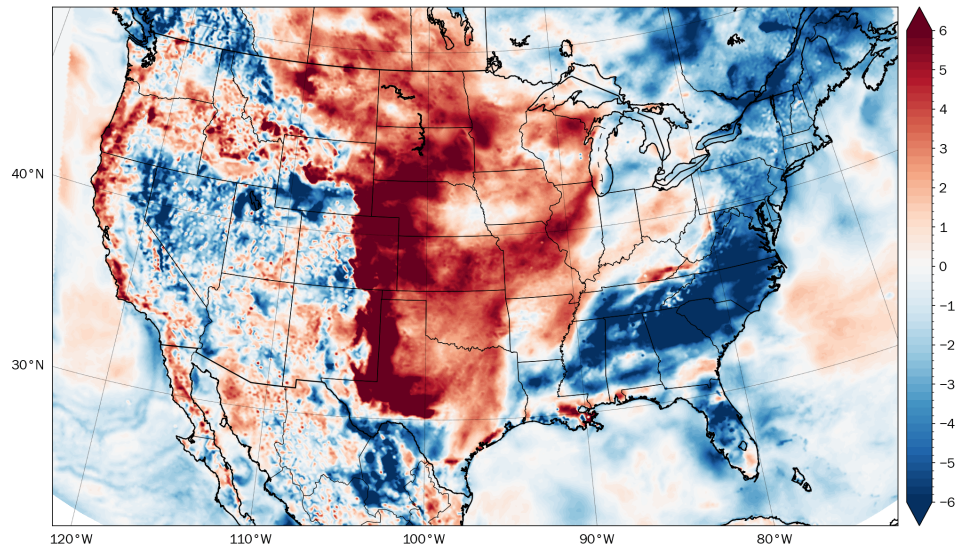


Figure 5.22: 2-meter temperature bias ($^{\circ}\text{C}$) from the MPAS control simulation relative to RTMA at forecast hour 66, valid at 1800 UTC 10 February 2021.

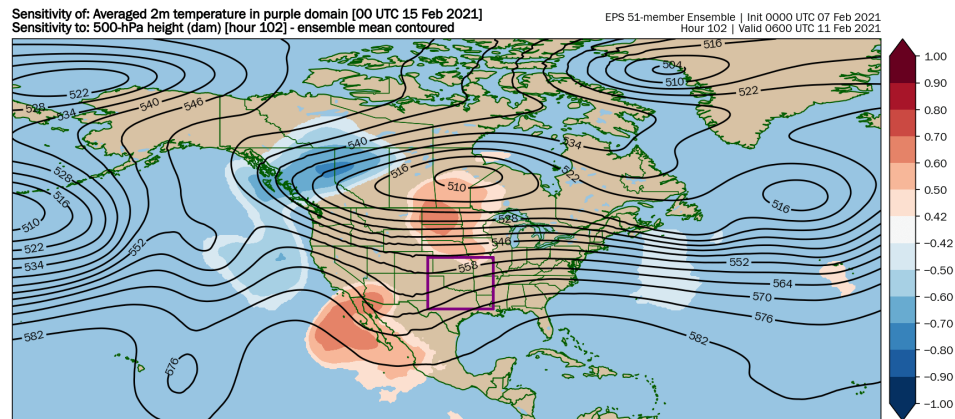


Figure 5.23: Sensitivity of 2-meter temperatures averaged over the southern Great Plains at 0000 UTC 15 February 2021 to 500-hPa geopotential height valid 0600 UTC 11 February 2021.

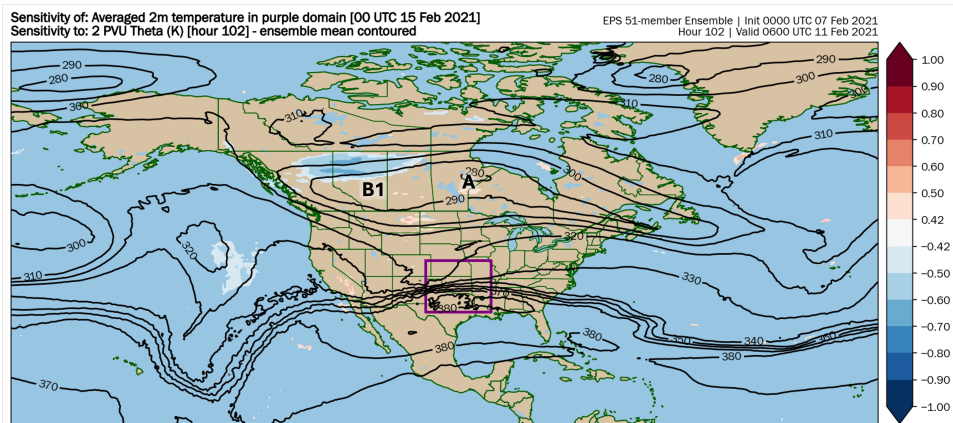


Figure 5.24: Same as Fig. 5.23, but showing sensitivity to 2 PVU potential temperature.

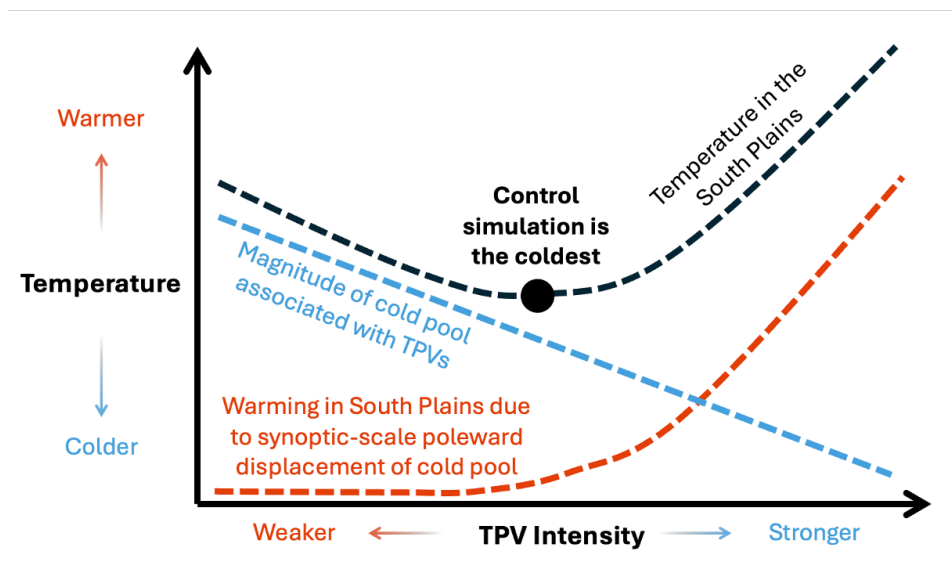


Figure 5.25: Schematic showing the general relationship between TPV intensity and cold pool intensity (blue line) contrasting with a different evolution of the PV lobe over western Canada in simulations with a more intense TPV than the control simulation leading to a poleward displacement of the core of the cold pool (red line). The end result is the control simulation producing the coldest temperatures in the southern Great Plains (black line).

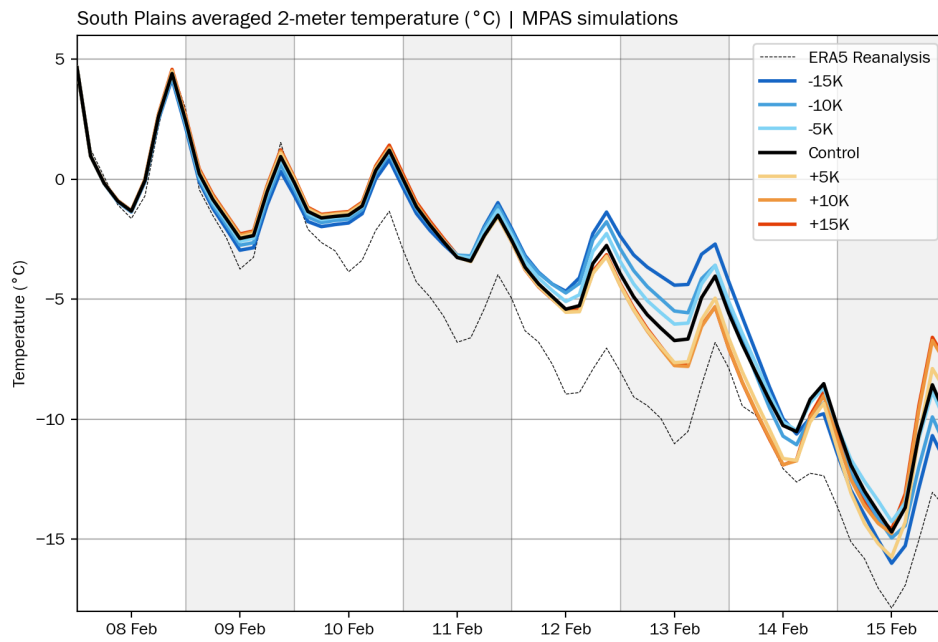


Figure 5.26: Same as Fig. 5.19, but for the MPAS experiments that did not incorporate a moving domain for the TPV modification region that filters out TPV B2 and a PV lobe over southwest Canada.

Chapter 6

Conclusions

The overarching goal of this dissertation was to advance our understanding of TPV linkages to rapid deepening of high-latitude cyclones, such as ACs and PLs, and TPV linkages to midlatitude cold air outbreaks. Specifically, the primary objectives were to: 1) assess characteristics of rapidly deepening ACs and their associations with TPVs, 2) examine and compare intense AC and PL cases and the role of TPV intensity in their track and intensity evolution from a numerical modeling framework, and 3) investigate the role of two merging TPVs in a historic CAO in the southern Great Plains.

Chapter 3 evaluated systematic characteristics of rapidly deepening ACs and their associated EGR, proximity to the nearest TPV, spatial distributions, and QG forcing for ascent. The hypotheses for this chapter were as follows:

- A1: Summertime RD cases are associated with stronger forcing from a nearby TPV, longer lifetimes, and closer proximity of the nearest TPV during the RD episode relative to wintertime cases.
- A2: RD cases along the poleward extension of the North Atlantic climatological baroclinic zone exhibit longer lifetimes before the RD episode, have a greater relative importance of lower-tropospheric baroclinic forcing, are more favored during the winter, and are farther from the nearest TPV during the rapid deepening episode relative to Arctic cases.
- A3: Cases that begin RD from the onset of when the cyclone is first trackable are more favored to be in the Arctic basin and are closer to the nearest TPV during their onset, relative to cases that exist at least 24 h prior to the onset of RD.

- A4: Processes associated with RD ACs are not inherently different than ordinary deepening ACs, but are associated with larger magnitudes of differential cyclonic vorticity advection and lower-tropospheric baroclinic instability, with constructive interference resulting in greater cyclone intensification rates, and are more likely to be associated with a nearby TPV during their peak growth rate than ordinary deepening ACs. Ordinary ACs for this purpose are defined as ACs that deepen at least 4 hPa over a 24 h period, but fail to meet RD criteria.

Rapidly deepening ACs were selected based on criteria normalizing cyclone deepening rates by a latitudinal climatology of longitudinally averaged observed cyclone deepening rates. Spatial composites rotated in the direction of the mean 250-850 hPa wind vector showed rapidly deepening ACs are associated with a composite mean TPV upstream and left of the AC at the beginning of the rapid deepening episode, with the TPV becoming vertically aligned with the surface cyclone by the end of the rapid deepening episode as a PV hook signature becomes evident in the composite mean. There was a weak composite mean signature for an antecedent AC downstream and left of the developing AC, which merges into the newly dominant AC as it rapidly deepens.

Summer RD cases were associated with a closer proximity to the closest TPV, lower EGR and EGR normalized anomalies, and overall longer lifetimes, relative to winter AC cases. The spatial distribution of summer RD cases was more heavily concentrated in the open Arctic and the Siberian and North American coastlines, while winter RD cases were more heavily concentrated near the North Atlantic region. These results confirm hypothesis A1. Partitioning RD cases by basin showed Atlantic cases are more frequent during the winter than summer months, are associated with more positive EGR but less positive EGR normalized anomaly relative to Pacific cases, and do not have significantly different distance from the closest TPV relative to summer cases. These results fail to confirm A2, and suggest partitioning cases by basin may not be an adequate method of

partitioning physically meaningful differences in characteristics of RD cases especially given the subjective nature of basin partitioning.

ACs whose RD onset coincides with cyclogenesis were less favored in the central Arctic basin, and are more distant from the closest TPV during the RD onset, relative to ACs which are trackable at least 24 h or more prior to the onset of RD. This result fails to confirm A3, although there are possible caveats to this result such as the sensitivity of the time of AC genesis to the choice of cyclone identification and tracking algorithm in addition to the dataset used. Lastly, comparing RD ACs to ordinary ACs that intensify but fail to meet RD criteria do not show a significant difference in the distance to the closest TPV, but RD cases are associated with greater EGR and EGR normalized anomaly relative to ordinary ACs. This result partially supports A4. Using QG forcing may not be fully adequate to analyze TPV-AC linkages, as the spatial smoothing applied to QG forcing terms is adequate for synoptic-scale troughs but may result in too much smoothing of sub-synoptic scale TPVs.

The results from Chapter 3 generally support the expectation that the spatial climatology and characteristics of the most rapidly deepening ACs are similar to the peak deepening rate of the most intense ACs (e.g., Gray et al. 2021; Vessey et al. 2022). A noticeable difference in this analysis compared to these studies is a more distinct PV hook signature in the composite mean 2 PVU field, which is in part due to the different methodology of rotating data relative to the tropospheric-mean wind vector as opposed to the cyclone motion vector. This method of rotating data into dynamically relevant coordinates, in addition to using TPV basins to match TPVs to ACs as opposed to the coordinate of the minimum 2 PVU potential temperature, can be adapted for use in future TPV-AC linkage studies.

Chapter 4 evaluated two case studies of high-latitude rapidly deepening cyclones associated with TPVs, focusing on the 2012 great Arctic cyclone (AC12) and a February 1987 polar low in Norway (PL87). The hypotheses for this chapter were as follows:

- B1: A more intense TPV results in a greater peak deepening rate, stronger peak intensity, and an earlier peak intensity of the surface cyclone relative to the control simulation, with the opposite true for a weaker TPV.
- B2: For the AC case, having two nearby TPVs being more intense results in a greater peak intensity of the AC relative to the control simulation or intensifying each TPV individually.
- B3: For the PL case, a weaker TPV results in a delayed occlusion of the surface cyclone and transition to an axisymmetric warm core, and a weaker cyclone overall, relative to the control simulation.

The AC12 experiment found the intensity and track of the cyclone were sensitive to the intensity of two TPVs, with the intensity of TPV1 primarily affecting the intensity of AC12, and the intensity of TPV2 primarily affecting the track of AC12. The impact of stronger TPVs on the peak intensity of AC12 was less than their impact on the peak deepening rate and timing of peak intensity of AC12, with stronger TPVs resulting in a faster deepening rate and earlier maturation and occlusion of AC12 than weaker TPVs. This result was predominantly associated with more positive differential cyclonic vorticity advection over the cyclone center in simulations with more intense TPVs. On the contrary, weakening the broader upper-level trough TPV1 is embedded in resulted in a greater shift of the cyclone track and intensity relative to intensifying the smaller-scale TPVs. Weakening the upper-level trough resulted in a weaker and more progressive cyclone, with a weaker peak deepening rate and a delayed peak intensity, relative to the control simulation. These results confirm both B1 and B2.

The PL87 experiment found strong sensitivity to the simulation ICs, with a pronounced inflection point in the modeled evolution of the cyclone at 1200 UTC 25 February 1987. Simulations initialized before this inflection point, even at only 18–24 h forecast lead time before the peak intensity of the PL, failed to simulate the observed PL genesis through

rapid cyclogenesis and warm seclusion and southward turn towards Norway. Simulations initialized after the inflection point, however, correctly resolved this observed evolution. An ensemble of simulations before and after the inflection point initialized from the ERA5 IC uncertainty ensemble failed to encompass any members that crossed the inflection point, suggesting that variability within the IC uncertainty ensemble fails to encompass the source of the inflection point in the ICs. Intensifying the TPV prior to the inflection point resulted in only a marginal shift towards a slower and more intense cyclone, while weakening the TPV after the inflection point resulted in only a marginal shift towards a weaker and more progressive cyclone. While these results are broadly in line with the expectation from B3, they are too small in magnitude to be physically meaningful.

The results of the AC12 experiment add a new TPV intensity perspective to the numerous existing studies on the evolution of AC12, showing that stronger TPVs are associated with a more rapid deepening rate and an earlier occlusion and peak intensity of AC12. This study further complements Chen et al. (2022) by focusing on the impact of a second upstream TPV on the evolution of AC12, finding a similar result in that cold air advection from the cold pool of TPV1 initially limited the deepening of AC12 in spite of differential cyclonic vorticity advection associated with TPV1. A new result is that the intensity of TPV2 predominantly affected the steering of AC12 at its peak deepening rate through modifying the amplitude of the downstream ridge and tropospheric-mean steering flow, while the intensity of TPV1 predominantly affected the deepening rate of AC12 through affecting the magnitude of differential cyclonic vorticity advection.

The results of the PL87 experiment, meanwhile, tied into the existing literature finding poor short-term forecast skill of PL development, even with high-resolution simulations (e.g., McInnes et al. 2011; Stoll et al. 2020; Doyle and Biernat 2023). While it is known that upper-tropospheric PV anomalies are on occasion necessary precursors to PL development (e.g., Mallet et al. 2013), literature on PL linkages to TPVs is limited, and this study did not add any definitive evidence as to whether TPV intensity, or the presence of a TPV, plays an

important role in PL development. However, ESA performed for this experiment showed the importance of environmental factors affecting the symmetry and gradient of the TPV, which in turn affected the development of the baroclinic low that subsequently developed into PL87, and these results further highlighted the importance of IC quality to correctly predicting PL evolution.

Altogether, the results of the PL87 and AC12 case studies show how TPV impacts on surface cyclone intensity and track can differ for two different phenomena. A simple spatial proximity of a surface cyclone to a nearby TPV does not necessarily imply that the intensity of the TPV has a significant impact on the evolution of the surface cyclone, which is likely more of the case for PLs than ACs given the limited existing literature on TPV-PL linkages. A takeaway point from this chapter is that the impact of TPV intensity on the evolution of surface cyclones is of a smaller magnitude than that of synoptic-scale upper tropospheric troughs. A possible explanation is that as TPVs are sub-synoptic scale features, their associated forcing for ascent is on a smaller spatial scale and results in less modulation of the downstream waveguide than that associated with larger wavelength troughs.

Chapter 5 evaluated the February 2021 CAO in the Southern Great Plains, focusing on the impact of two merging TPVs over southern Canada on the evolution of the CAO. The hypotheses for this chapter were as follows:

- C1: Stronger TPVs are expected to merge at a faster rate than weaker TPVs.
- C2: Stronger TPVs are expected to lead to a stronger cold outbreak in the southern Great Plains than weaker TPVs.
- C3: The presence of multiple TPVs and their associated merger prolonged the duration of the CAO in the southern Great Plains, as a faster merger would be associated with the merged TPV propagating downstream earlier, and accordingly a shorter duration of the CAO in the southern Great Plains, than a slower merger.

A synoptic overview of the CAO reveals its anomalous duration and magnitude was associated with a quasi-stationary large-scale upper-level trough, with two embedded TPVs gradually merging, with prolonged duration of shallow cold air damming across the southern Great Plains followed by more widespread cold air advection associated with the cold pool of the merged TPV. Backward air parcel trajectories originating from Oklahoma City at the peak of the CAO revealed parcels were originally constrained within the circulation of the TPVs before being ejected southward and descending in altitude as they reached Oklahoma. Regridding the parcel trajectories into TPV-relative coordinates showed parcels ending in Oklahoma City were more likely to be associated with TPV B1 as opposed to TPV A, despite the close proximity of both TPVs, suggesting that while the evolution of both TPVs was important for downstream impacts due to their merger, TPV B1 had a more direct role than TPV A in the CAO magnitude in the southern Great Plains.

The impact of TPV intensity on the peak magnitude of the CAO in the southern Great Plains was nonlinear, with the control simulation the coldest during the peak of the CAO. Simulations with weaker TPVs had a weaker cold pool and weaker cold air advection into the greater South Plains, while simulations with stronger TPVs had a stronger cold pool but with a farther north and more meridionally oriented resultant merged TPV, resulting in warmer temperatures in the southern Great Plains. This result fails to confirm C2, which stated that stronger TPVs would be associated with colder temperatures in the southern Great Plains. This result furthermore differs from a previous study of TPV intensity linkages to CAO magnitude by Lillo et al. (2021) which found a linear relationship between TPV intensity and CAO magnitude. This difference is likely due to the presence of a single, fast-moving TPV in their study compared to two slow-moving and merging TPVs in this study, as merging TPVs result in additional degrees of freedom accounting for their merger rate and impact on the ambient synoptic flow. More intense TPVs were found to merge at a faster rate than slower TPVs, thus confirming C1.

Hypothesis C3 stated that the TPV merger would prolong the duration of the CAO in the southern Great Plains. This hypothesis is not confirmed by this study, as simulations with a faster merger rate of TPVs ended up with a slower and more meridionally oriented resultant TPV than simulations with a slower merger rate. This result may be due to short-term variability in the eastern gradient of the TPVs quickly growing in scale at later forecast hours and modifying the synoptic-scale waveguide to result in a slower downstream progression of the upper-level trough.

There is a common theme throughout this dissertation relating to the role of TPVs in cyclogenesis and CAO evolution. The intensity of the TPV does have a role in cyclogenesis and CAO evolution through the magnitude of vorticity advection and the cold pool associated with the TPV. However, a more important source of forecast uncertainty is the gradient of the TPV, relating to its symmetry and environmental factors such as upstream ridges that can affect the symmetry and track of the TPV. Accordingly, even if a TPV is close to a developing surface cyclone, the intensity of the TPV may not significantly affect the track and intensity of the cyclone. Furthermore, the results of Chapter 5 showed that the presence of multiple interacting TPVs linked with a CAO result in nonlinear impacts on the spatial extent and magnitude of the CAO, compared to prior literature on single, fast-moving TPVs, which exhibit a more linear relationship between more intense TPVs being associated with a more expansive and colder CAO. It can further be hypothesized given these results that a CAO can occur without the presence of a TPV, but that the added presence of one or more TPVs embedded within a larger wavelength trough can increase the magnitude of the CAO if the TPVs are a source region of backward parcel trajectories from the location of the CAO.

6.1 Recommendations for future work

There are numerous outstanding questions remaining from the results of this dissertation, which in addition to the conclusions previously presented, present numerous opportunities for future work.

Results from Chapter 3 focused on the first RD episode of ACs. However, it was also found that 20 AC cases in the time period of interest (2.4% of all RD cases) underwent 2 or more RD episodes. Given the well-known fact that ACs can be exceptionally long-lived, especially in the summer, and that about 48% of ACs annually are associated with at least two or more TPVs (Cavallo and Burg 2024), it can be speculated that ACs with multiple RD episodes are more likely in the summer, and are more likely to be associated with multiple TPVs, than ACs with a single RD episode. As such, it may be of interest to further investigate ACs that undergo multiple RD episodes, including their linkages to multiple TPV interactions. A related open question is whether the presence of multiple RD episodes for a single AC is an artifact of a cyclone identification and tracking algorithm, especially for multi-center cyclones, or whether it is unambiguously the same AC as identified by a continuous, single center in MSLP and low-level vorticity fields.

While the cases evaluated in Chapter 4 were predominantly from a synoptic perspective, maturing cyclones can be associated with mesoscale impacts, such as intense snow bands in the cold sector and severe thunderstorms in the warm sector. Slight changes in the amplitude and track of the surface cyclone on the order of those seen in the results of chapter 4 can have an outsized societal impact if the gradient of potential heavy snowfall or severe weather overlaps large population centers. From an Arctic perspective, none of the simulations performed in this dissertation included coupling with variable sea ice, though given the rapid sea ice loss associated with AC12 as it stalled over a region of marginal sea ice, an outstanding question is whether track and intensity changes on the order of those resulting from the TPV intensity variability imposed can significantly affect the rate of sea ice loss, or if the sea ice edge in turn affects the track and intensification rate of the AC.

Results from Chapter 5 provide multiple opportunities for future work, including on research and operational forecast tools. A unique aspect of the February 2021 CAO is the presence of multiple merging TPVs. The methodology used for converting parcel trajectories into TPV-centered coordinates can be applied in addition to ESA to identify whether any one TPV has an outsized role in a CAO evolution for a region of interest, which may allow forecasters and researchers alike to target more analysis efforts on one TPV over the other. A related research opportunity is to repeat the TPV modification experiments for each TPV individually, which can additionally quantify if modifying the intensity of TPV B1 would have a greater impact on the CAO evolution in the Southern Great Plains than modifying TPV A.

Bibliography

- Albers, J. R., M. Newman, A. Hoell, M. L. Breeden, Y. Wang, and J. Lou, 2022: The February 2021 cold air outbreak in the United States: A subseasonal forecast of opportunity. *Bull. Amer. Meteor. Soc.*, **103** (12), E2887–E2904, doi:10.1175/BAMS-D-21-0266.1.
- Allen, J. T., A. B. Pezza, and M. T. Black, 2010: Explosive cyclogenesis: A global climatology comparing multiple reanalyses. *J. Climate*, **23**, 6468–6484, doi:10.1175/2010JCLI3437.1.
- Ancell, B. C., A. Bogusz, M. J. Lauridsen, and C. J. Nauert, 2018: Seeding Chaos: The dire consequences of numerical noise in NWP perturbation experiments. *Bull. Amer. Meteor. Soc.*, **99**, 615–628, doi:10.1175/BAMS-D-17-0129.1.
- Ancell, B. C. and A. A. Coleman, 2022: New perspectives on ensemble sensitivity analysis with applications to a climatology of severe convection. *Bull. Amer. Meteor. Soc.*, **103**, E511–E530, doi:10.1175/BAMS-D-20-0321.1.
- Ancell, B. C. and G. J. Hakim, 2007: Comparing adjoint- and ensemble-sensitivity analysis with applications to observation targeting. *Mon. Wea. Rev.*, **135**, 4117–4134, doi:10.1175/2007MWR1904.1.
- Aspelien, T., T. Iversen, J. B. Bremnes, and I. L. Frogner, 2011: Short-range probabilistic forecasts from the Norwegian limited-area EPS: Long-term validation and a polar low study. *Tellus A*, **63**, 564–584, doi:10.1111/j.1600-0870.2010.00502.x.
- Barnhart, K. R., I. Overeem, and R. S. Anderson, 2014: The effect of changing sea ice on the physical vulnerability of Arctic coasts. *Cryosphere*, **8**, 1777–1799, doi:10.5194/tc-8-1777-2014.
- Bell, G. D. and L. F. Bosart, 1989: A 15-year climatology of Northern Hemisphere 500 mb closed cyclone and anticyclone centers. *Mon. Wea. Rev.*, **117**, 2142–2163, doi:10.1175/1520-0493(1989)117<2142:AYCONH>2.0.CO;2.
- Bentley, A. M., 2018: Extratropical cyclones leading to extreme weather events over central and eastern north america. Phd thesis, University at Albany.
- Berman, J. D. and R. D. Torn, 2019: The impact of initial condition and warm conveyor belt forecast uncertainty on variability in the downstream waveguide in an ECWMF case study. *Mon. Wea. Rev.*, **147**, 4071–4089, doi:10.1175/MWR-D-18-0333.1.
- Bethan, S., G. Vaughan, and S. J. Reid, 1996: A comparison of ozone and thermal tropopause heights and the impact of tropopause definition on quantifying the ozone content of the troposphere. *Quart. J. Roy. Meteor. Soc.*, **122**, 929–944, doi:10.1002/qj.49712253207.

- Biernat, K. A., L. F. Bosart, and D. Keyser, 2021: A climatological analysis of the linkages between tropopause polar vortices, cold pools, and cold air outbreaks over the central and eastern United States. *Mon. Wea. Rev.*, **149**, 189–206, doi:10.1175/MWR-D-20-0191.1.
- Biernat, K. A., D. Keyser, and L. F. Bosart, 2023: A climatological comparison of the Arctic environment and Arctic cyclones between periods of low and high forecast skill of the synoptic-scale flow. *Mon. Wea. Rev.*, **151**, 1957–1978, doi:10.1175/MWR-D-22-0318.1.
- Blanchard-Wrigglesworth, E., M. Webster, L. Boisvert, C. Parker, and C. Horvat, 2022: Record arctic cyclone of January 2022: characteristics, impacts, and predictability. *J. Geophys. Res. Atmos.*, **127**, e2022JD037161, doi:10.1029/2022JD037161.
- Bluestein, H. B., 1992: *Principles of Kinematics and Dynamics. Vol. I, Synoptic-Dynamic Meteorology in Midlatitudes*. Oxford University Press, 431 pp.
- Bolinger, R. A., et al., 2022: An assessment of the extremes and impacts of the February 2021 South-Central U.S. Arctic outbreak, and how climate services can help. *Weather and Climate Extremes*, **36** (April), 100461, doi:10.1016/j.wace.2022.100461.
- Borg, S. M., S. M. Cavallo, and D. D. Turner, 2020: Characteristics of tropopause polar vortices based on observations over the Greenland ice sheet. *J. Appl. Meteor. Climatol.*, **59**, 1933–1947, doi:10.1175/JAMC-D-20-0004.1.
- Bosart, L. F., 1981: The Presidents’ Day Snowstorm of 18–19 February 1979: A subsynoptic-scale event. *Mon. Wea. Rev.*, **109**, 1542–1566, doi:10.1175/1520-0493(1981)109<1542:TPDSOF>2.0.CO;2.
- Bosart, L. F., G. J. Hakim, K. R. Tyle, M. A. Bedrick, E. W. Bracken, M. J. Dickinson, and D. M. Schultz, 1996: Large-scale antecedent conditions associated with the 12–14 March 1993 cyclone (“Superstorm ’93”) over eastern North America. *Mon. Wea. Rev.*, **124**, 1865–1891, doi:10.1175/1520-0493(1996)124<1865:LSACAW>2.0.CO;2.
- Bosart, L. F. and S. C. Lin, 1984: A diagnostic analysis of the Presidents’ Day Storm of February 1979. *Mon. Wea. Rev.*, **112**, 2148–2177, doi:10.1175/1520-0493(1984)112<2148:ADAOTP>2.0.CO;2.
- Boyle, J. S. and L. F. Bosart, 1983: A cyclone/anticyclone couplet over North America: An example of anticyclone evolution. *Mon. Wea. Rev.*, **111**, 1025–1045, doi:10.1175/1520-0493(1983)111<1025:ACCONA>2.0.CO;2.
- Boyle, J. S. and L. F. Bosart, 1986: Cyclone-anticyclone couplets over North America. Part II: Analysis of a major cyclone event over the eastern United States. *Mon. Wea. Rev.*, **114**, 2432–2465, doi:10.1175/1520-0493(1986)114<2432:CCONAP>2.0.CO;2.
- Bracegirdle, T. J. and S. L. Gray, 2009: The dynamics of a polar low assessed using potential vorticity inversion. *Quart. J. Roy. Meteor. Soc.*, **135**, 880–893, doi:10.1002/qj.411.

- Bray, M. T., S. M. Cavallo, and H. B. Bluestein, 2021: Examining the relationship between tropopause polar vortices and tornado outbreaks. *Wea. Forecasting*, **36**, 1799–1814, doi:10.1175/WAF-D-21-0058.1.
- Businger, S. and R. J. Reed, 1989: Cyclogenesis in cold air masses. *Wea. Forecasting*, **4**, 133–156.
- Butler, A. H., J. P. Sjöberg, D. J. Seidel, and K. H. Rosenlof, 2017: A sudden stratospheric warming compendium. *Earth Syst. Sci. Data*, **9**, 63–76, doi:10.5194/essd-9-63-2017.
- Capute, P. K. and R. D. Torn, 2021: A comparison of Arctic and Atlantic cyclone predictability. *Mon. Wea. Rev.*, **149**, 3837–3849, doi:10.1175/MWR-D-20-0350.1.
- Cavallo, S. M. and T. Burg, 2024: The impacts of tropopause vortex-vortex interactions on arctic cyclones. *Polar Meteorology and Oceanography Special Symposium*, Baltimore, MD, Amer. Meteor. Soc.
- Cavallo, S. M. and G. J. Hakim, 2009: Potential vorticity diagnosis of a tropopause polar cyclone. *Mon. Wea. Rev.*, **137**, 1358–1371, doi:10.1175/2008MWR2670.1.
- Cavallo, S. M. and G. J. Hakim, 2010: Composite structure of tropopause polar cyclones. *Mon. Wea. Rev.*, **138**, 3840–3857, doi:10.1175/2010MWR3371.1.
- Cavallo, S. M. and G. J. Hakim, 2012: Radiative impact on tropopause polar vortices over the Arctic. *Mon. Wea. Rev.*, **140**, 1683–1702, doi:10.1175/MWR-D-11-00182.1.
- Cavallo, S. M. and G. J. Hakim, 2013: Physical mechanisms of tropopause polar vortex intensity change. *J. Atmos. Sci.*, **70**, 3359–3373, doi:10.1175/JAS-D-13-088.1.
- Chen, L. J., G. Fu, and P. Y. Li, 2022: A new perspective on the development of the great Arctic cyclone in August 2012. *J. Geophys. Res. Atmos.*, **127**, e2022JD036818, doi:10.1029/2022JD036818.
- Clancy, R., C. M. Bitz, B.-W. Edward, M. C. McGraw, and S. M. Cavallo, 2022: A cyclone-centered perspective on the drivers of asymmetric patterns in the atmosphere and sea ice during Arctic cyclones. *J. Climate*, **35**, 73–89, doi:10.1175/JCLI-D-21-0093.1.
- Claud, C., G. Heinemann, E. Raustein, and L. McMurdie, 2004: Polar low le Cygne: Satellite observations and numerical simulations. *Quart. J. Roy. Meteor. Soc.*, **130** (**598 PART A**), 1075–1102, doi:10.1256/qj.03.72.
- Coleman, A. A. and B. C. Ancell, 2020: Toward the improvement of high-impact probabilistic forecasts with a sensitivity-based convective-scale ensemble subsetting technique. *Mon. Wea. Rev.*, **148**, 4995–5014, doi:10.1175/MWR-D-20-0043.1.
- Crawford, A. D., E. A. Schreiber, N. Sommer, M. C. Serreze, J. C. Stroeve, and D. G. Barber, 2021: Sensitivity of northern hemisphere cyclone detection and tracking results to fine spatial and temporal resolution using ERA5. *Mon. Wea. Rev.*, **149**, 2581–2598, doi:10.1175/MWR-D-20-0417.1.

- Crawford, A. D. and M. C. Serreze, 2016: Does the summer Arctic frontal zone influence Arctic ocean cyclone activity? *J. Climate*, **29**, 4977–4993, doi:10.1175/JCLI-D-15-0755.1.
- Croad, H. L., J. Methven, B. Harvey, S. P. Keeley, A. Volonté, and K. I. Hodges, 2023: A climatology of summer-time Arctic cyclones using a modified phase space. *Geophys. Res. Lett.*, **50**, 1–10, doi:10.1029/2023GL105993.
- Curry, J., W. B. Rossow, D. Randall, and J. L. Schramm, 1996: Overview of Arctic cloud and radiation characteristics. *J. Climate*, **9**, 1731–1764, doi:10.1175/1520-0442(1996)009<1731:OOACAR>2.0.CO;2.
- Davis, C. A. and K. A. Emanuel, 1991: Potential vorticity diagnostics of cyclogenesis. *Mon. Wea. Rev.*, **119**, 1929–1953, doi:10.1175/1520-0493(1991)119<1929:PVDOC>2.0.CO;2.
- Davis, N. A., J. H. Richter, A. A. Glanville, J. Edwards, and E. LaJoie, 2022: Limited surface impacts of the January 2021 sudden stratospheric warming. *Nature Communications*, **13**, 1–11, doi:10.1038/s41467-022-28836-1.
- Day, J. J., I. Sandu, L. Magnusson, M. J. Rodwell, H. Lawrence, N. Bormann, and T. Jung, 2019: Increased Arctic influence on the midlatitude flow during Scandinavian blocking episodes. *Quart. J. Roy. Meteor. Soc.*, **145**, 3846–3862, doi:10.1002/qj.3673.
- De Pondeca, M. S., et al., 2011: The Real-Time Mesoscale Analysis at NOAA’s National Centers for Environmental Prediction: Current status and development. *Wea. Forecasting*, **26**, 593–612, doi:10.1175/WAF-D-10-05037.1.
- Domeisen, D. I. and A. H. Butler, 2020: Stratospheric drivers of extreme events at the Earth’s surface. *Commun. Earth. Environ.*, **1**, 1–8, doi:10.1038/s43247-020-00060-z.
- Dowell, D. C., et al., 2022: The High-Resolution Rapid Refresh (HRRR): An hourly updating convection-allowing forecast model. Part I: Motivation and system description. *Wea. Forecasting*, **37**, 1371–1395, doi:10.1175/WAF-D-21-0151.1.
- Doyle, J. D. and K. Biernat, 2023: Sensitivity and predictability of mesoscale polar lows. *DRI meeting for seasonal prediction*, Monterey, CA, Naval Research Laboratory.
- Emanuel, K. A. and R. Rotunno, 1989: Polar lows as Arctic hurricanes. *Tellus A*, **41A**, 1–17, doi:10.1111/j.1600-0870.1989.tb00362.x.
- Fang, Z., P. T. Freeman, C. B. Field, and K. J. Mach, 2018: Reduced sea ice protection period increases storm exposure in Kivalina, Alaska. *Arctic Sci.*, **4**, 525–537, doi:10.1139/as-2017-0024.
- Finocchio, P. M., J. D. Doyle, and D. P. Stern, 2022: Accelerated sea ice loss from late summer cyclones in the new Arctic. *J. Climate*, **35** (23), 4151–4169, doi:10.1175/JCLI-D-22-0315.1.

- Finocchio, P. M., J. D. Doyle, D. P. Stern, and M. G. Fearon, 2020: Short-term impacts of Arctic summer cyclones on sea ice extent in the marginal ice zone. *Geophys. Res. Lett.*, **47** (13), 1–9, doi:10.1029/2020GL088338.
- Føre, I., J. E. Kristjánsson, Ø. Saetra, Ø. Breivik, B. Røsting, and M. Shapiro, 2011: The full life cycle of a polar low over the Norwegian Sea observed by three research aircraft flights. *Quart. J. Roy. Meteor. Soc.*, **137** (660), 1659–1673, doi:10.1002/qj.825.
- Gray, S. L., K. I. Hodges, J. L. Vautrey, and J. Methven, 2021: The role of tropopause polar vortices in the intensification of summer Arctic cyclones. *Wea. Climate Dynam.*, **2**, 1303–1324, doi:10.5194/wcd-2-1303-2021.
- Gyakum, J. R., J. R. Anderson, R. H. Grumm, and E. L. Gruner, 1989: North Pacific cold-season surface cyclone activity: 1975–1983. *Mon. Wea. Rev.*, **117**, 1141–1155, doi:10.1175/1520-0493(1989)117<1141:NPCCSSC>2.0.CO;2.
- Hakim, G. J., 2000: Climatology of coherent structures on the extratropical tropopause. *Mon. Wea. Rev.*, **128**, 385–406, doi:10.1175/1520-0493(2000)128<0385:COCSOT>2.0.CO;2.
- Hakim, G. J. and A. K. Canavan, 2005: Observed cyclone–anticyclone tropopause vortex asymmetries. *J. Atmos. Sci.*, **62**, 231–240, doi:10.1175/JAS-3353.1.
- Hakim, G. J., D. Keyser, and L. F. Bosart, 1996: The Ohio Valley wave-merger cyclogenesis event of 25–26 January 1978. Part II: Diagnosis using quasigeostrophic potential vorticity inversion. *Mon. Wea. Rev.*, **124**, 2176–2205, doi:10.1175/1520-0493(1996)124<2176:TOVWMC>2.0.CO;2.
- Hakim, G. J., C. Snyder, and D. J. Muraki, 2002: A new surface model for cyclone–anticyclone asymmetry. *J. Atmos. Sci.*, **59**, 2405–2420, doi:10.1175/1520-0469(2002)059<2405:ANSMFC>2.0.CO;2.
- Hartig, K., E. Tziperman, and C. P. Loughner, 2023: Processes contributing to North American cold air outbreaks based on air parcel trajectory analysis. *J. Climate*, **36**, 931–943, doi:10.1175/JCLI-D-22-0204.1.
- Heleniak, T., 2020: Polar peoples in the future: projections of the arctic populations. *Nordregio*, 1–62, doi:10.6027/WP2020:3.1403-2511.
- Hersbach, H., et al., 2020: The era5 global reanalysis. *Quart. J. Roy. Meteor. Soc.*, **146**, 1999–2049, doi:https://doi.org/10.1002/qj.3803.
- Hewson, T. D. and H. A. Titley, 2010: Objective identification, typing and tracking of the complete life-cycles of cyclonic features at high spatial resolution. *Meteor. Appl.*, **17**, 355–381, doi:10.1002/met.204.
- Hibbard, W., L. W. Uccellini, D. Santek, and K. F. Brill, 1989: Application of the 4-D McIDAS to a model diagnostic study of the Presidents’ Day cyclone. *Bull. Amer. Meteor. Soc.*, **70**, 1394–1403, doi:10.1175/1520-0477(1989)070<1394:AOTDMT>2.0.CO;2.

- Hodges, K. I., 1994: A general method for tracking analysis and its application to meteorological data. *Mon. Wea. Rev.*, **122**, 2573–2586, doi:10.1175/1520-0493(1994)122<2573:AGMFTA>2.0.CO;2.
- Holton, J. R., P. H. Haynes, M. E. McIntyre, A. R. Douglass, R. B. Rood, and L. Pfister, 1995: Stratosphere-troposphere exchange. *Rev. Geophys.*, **33**, 403–439, doi:10.1029/95RG02097.
- Hoskins, B. J., M. E. McIntyre, and A. W. Robertson, 1985: On the use and significance of isentropic potential vorticity maps. *Quart. J. Roy. Meteor. Soc.*, **111**, 877–946, doi:10.1002/qj.49711147002.
- Hoskins, P. J., B. J. Valdes, 1990: On the existence of storm-tracks. *J. Atmos. Sci.*, **47**, 1854–1864, doi:10.1175/1520-0469(1990)047<1854:OTEOST>2.0.CO;2.
- Ikeda, K., M. Steiner, and G. Thompson, 2017: Examination of mixed-phase precipitation forecasts from the high-resolution rapid refresh model using surface observations and sounding data. *Wea. Forecasting*, **32**, 949–967, doi:10.1175/WAF-D-16-0171.1.
- Intrieri, J. M., C. W. Fairall, M. D. Shupe, O. G. Persson, E. L. Andreas, P. S. Guest, and R. E. Moritz, 2002: An annual cycle of Arctic surface cloud forcing at SHEBA. *J. Geophys. Res.*, **107**, 8039, doi:10.1029/2000JC000439.
- Johnson, A. and X. Wang, 2021: Observation impact study of an Arctic cyclone associated with a tropopause polar vortex (TPV)-induced Rossby wave initiation event. *Mon. Wea. Rev.*, **149**, 1577–1591, doi:10.1175/MWR-D-20-0285.1.
- Kalnay, E. and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.*, **77**, 437–471, doi:10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2.
- Keyser, D. and M. A. Shapiro, 1986: A review of the structure and dynamics of upper-level frontal zones. *Mon. Wea. Rev.*, **114**, 452–499, doi:10.1175/1520-0493(1986)114<0452:AROTSA>2.0.CO;2.
- Kleinschmidt, E., 1950a: Über Aufbau und Entstehung von Zyklonen (1. Teil). *Meteor. Rundsch.*, **3**, 1–6.
- Kleinschmidt, E., 1950b: Über Aufbau und Entstehung von Zyklonen (2. Teil). *Meteor. Rundsch.*, **3**, 54–61.
- Kleinschmidt, E., 1951: Über Aufbau und Entstehung von Zyklonen (3. Teil). *Meteor. Rundsch.*, **4**, 89–96.
- Kleinschmidt, E., 1955: Die Entstehung einer Höhenzyklone über Nordamerika. *Tellus*, **7**, 96–110.
- Kleinschmidt, E., 1957: Cyclones and anticyclones. *Dynamic Meteorology*, A. Eliassen and E. Kleinschmidt, Eds., Handbuch der Physik, Vol. 48, 112–129.

- Knapp, K. R., M. C. Kruk, D. H. Levinson, H. J. Diamond, and C. J. Neumann, 2010: The international best track archive for climate stewardship (IBTrACS). *Bull. Amer. Meteor. Soc.*, **91**, 363–376, doi:10.1175/2009BAMS2755.1.
- Kolstad, E. W. and T. J. Bracegirdle, 2017: Sensitivity of an apparently hurricane-like polar low to sea-surface temperature. *Quart. J. Roy. Meteor. Soc.*, **143**, 966–973, doi:10.1002/qj.2980.
- Kolstad, E. W., T. Breiteig, and A. A. Scaife, 2010: The association between stratospheric weak polar vortex events and cold air outbreaks in the Northern Hemisphere. *Quart. J. Roy. Meteor. Soc.*, **136**, 886–893, doi:10.1002/qj.620.
- Lee, S. H., J. C. Furtado, and A. J. Charlton-Perez, 2019: Wintertime North American weather regimes and the Arctic Stratospheric polar vortex. *Geophys. Res. Lett.*, **46**, 14 892–14 900, doi:10.1029/2019GL085592.
- Li, X. and A. H. Lynch, 2023: New insights into projected Arctic sea road: operational risks, economic values, and policy implications. *Climatic Change*, **176** (4), 30, doi:10.1007/s10584-023-03505-4.
- Lillo, S. P., S. M. Cavallo, D. B. Parsons, and C. Riedel, 2021: The role of a tropopause polar vortex in the generation of the January 2019 extreme Arctic outbreak. *J. Atmos. Sci.*, **78**, 2801–2821, doi:10.1175/JAS-D-20-0285.1.
- Lim, E. P. and I. Simmonds, 2002: Explosive cyclone development in the Southern Hemisphere and a comparison with Northern Hemisphere events. *Mon. Wea. Rev.*, **130**, 2188–2209, doi:10.1175/1520-0493(2002)130<2188:ECDITS>2.0.CO;2.
- Mallet, P. E., C. Claud, C. Cassou, G. Noer, and K. Kodera, 2013: Polar lows over the nordic and labrador seas: Synoptic circulation patterns and associations with north atlantic-europe wintertime weather regimes. *J. Geophys. Res. Atmos.*, **118** (6), 2455–2472, doi:10.1002/jgrd.50246.
- Martin, R. and G. W. Moore, 2006: Transition of a synoptic system to a polar low via interaction with the orography of Greenland. *Tellus, Series A: Dynamic Meteorology and Oceanography*, **58** (2), 236–253, doi:10.1111/j.1600-0870.2005.00169.x.
- May, R. M., et al., 2022: Metpy: A meteorological python library for data analysis and visualization. *Bull. Amer. Meteor. Soc.*, **103**, E2273–E2284, doi:10.1175/BAMS-D-21-0125.1.
- McInnes, H., J. Kristiansen, J. E. Kristjánsson, and H. Schyberg, 2011: The role of horizontal resolution for polar low simulations. *Quart. J. Roy. Meteor. Soc.*, **137** (660), 1674–1687, doi:10.1002/qj.849.
- Merryfield, W. J., et al., 2020: Current and emerging developments in subseasonal to decadal prediction. *Bull. Amer. Meteor. Soc.*, **101**, E869–E896, doi:10.1175/BAMS-D-19-0037.1.

- Millin, O. T. and J. C. Furtado, 2022: The role of wave breaking in the development and subseasonal forecasts of the February 2021 Great Plains cold air outbreak. *Geophys. Res. Lett.*, **49** (21), doi:10.1029/2022GL100835.
- Millin, O. T., J. C. Furtado, and J. B. Basara, 2022: Characteristics, evolution, and formation of cold air outbreaks in the Great Plains of the United States. *J. Climate*, **35**, 4585–4602, doi:10.1175/JCLI-D-21-0772.1.
- Minder, J. R., et al., 2023: P-type processes and predictability. *Bull. Amer. Meteor. Soc.*, **104**, E1469–E1492, doi:https://doi.org/10.1175/BAMS-D-22-0095.1.
- Montgomery, M. T. and B. F. Farrell, 1992: Polar low dynamics. *J. Atmos. Sci.*, **49**, 2484–2505, doi:10.1175/1520-0469(1992)049<2484:PLD>2.0.CO;2.
- Moore, G. W., M. C. Reader, J. York, and S. Sathiyamoorthy, 1996: Polar lows in the Labrador Sea: A case study. *Tellus, Series A: Dynamic Meteorology and Oceanography*, **48A**, 17–40, doi:10.1034/j.1600-0870.1996.00002.x.
- Moreno-Ibáñez, M., R. Laprise, and P. Gachon, 2021: Recent advances in polar low research: current knowledge, challenges and future perspectives. *Tellus, Series A: Dynamic Meteorology and Oceanography*, **73** (1), 1–31, doi:10.1080/16000870.2021.1890412, URL <https://doi.org/10.1080/16000870.2021.1890412>.
- Morgan, M. C. and J. W. Nielsen-Gammon, 1998: Using tropopause maps to diagnose midlatitude weather systems. *Mon. Wea. Rev.*, **126**, 2555–2579, doi:10.1175/1520-0493(1998)126<2555:UTMTDM>2.0.CO;2.
- Mudryk, L. R., J. Dawson, S. E. Howell, C. Derksen, T. A. Zagon, and M. Brady, 2021: Impact of 1, 2 and 4 C of global warming on ship navigation in the Canadian Arctic. *Nat. Climate Change*, **11**, 673–679, doi:10.1038/s41558-021-01087-6.
- Neu, U., et al., 2013: IMILAST: A community effort to intercompare extratropical cyclone detection and tracking algorithms. *Bull. Amer. Meteor. Soc.*, **94**, 529–547, doi:10.1175/BAMS-D-11-00154.1.
- Nielsen, D. M., P. Pieper, A. Barkhordarian, P. Overduin, T. Ilyina, V. Brovkin, J. Baehr, and M. Dobrynin, 2022: Increase in Arctic coastal erosion and its sensitivity to warming in the twenty-first century. *Nat. Climate Change*, **12** (3), 263–270, doi:10.1038/s41558-022-01281-0.
- Papritz, L., E. Rouges, F. Aemisegger, and H. Wernli, 2019: On the thermodynamic preconditioning of Arctic air masses and the role of tropopause polar vortices for cold air outbreaks from Fram Strait. *J. Geophys. Res. Atmos.*, **124**, 33–50, doi:10.1029/2019JD030570.
- Park, S. H., J. B. Klemp, and W. C. Skamarock, 2014: A comparison of mesh refinement in the global MPAS-A and WRF models using an idealized normal-mode baroclinic wave simulation. *Mon. Wea. Rev.*, **142** (10), 3614–3634, doi:10.1175/MWR-D-14-00004.1.

- Pedlosky, J., 1998: *Geophysical Fluid Dynamics*. 2d ed., Springer, 728 pp.
- Petrack, S., K. Riemann-Campe, S. Hoog, C. Growitsch, H. Schwind, R. Gerdes, and K. Rehdanz, 2017: Climate change, future Arctic sea ice, and the competitiveness of European Arctic offshore oil and gas production on world markets. *Ambio*, **46**, 410–422, doi:10.1007/s13280-017-0957-z.
- Pyle, M. E., D. Keyser, and L. F. Bosart, 2004: A diagnostic study of jet streaks: Kinematic signatures and relationship to coherent tropopause disturbances. *Mon. Wea. Rev.*, **132**, 297–319, doi:10.1175/1520-0493(2004)132<0297:ADSOJS>2.0.CO;2.
- Reynolds, C. A., J. D. Doyle, F. M. Ralph, and R. Demirdjian, 2019: Adjoint sensitivity of North Pacific atmospheric river forecasts. *Mon. Wea. Rev.*, **147**, 1871–1897, doi:10.1175/MWR-D-18-0347.1.
- Riedel, C., 2020: Tropospheric polar vortices and impacts on atmospheric flow from the arctic to the mid-latitudes using a new global modeling system. Phd thesis, University of Oklahoma.
- Rios-Berrios, R., R. D. Torn, and C. A. Davis, 2016: An ensemble approach to investigate tropical cyclone intensification in sheared environments. Part I: Katia (2011). *J. Atmos. Sci.*, **73**, 71–93, doi:10.1175/JAS-D-15-0052.1.
- Roebber, P. J., 1984: Statistical analysis and updated climatology of explosive cyclones. *Mon. Wea. Rev.*, **112**, 1577–1589, doi:10.1175/1520-0493(1984)112<1577:SAAUO>2.0.CO;2.
- Sanders, F., 1986: Explosive cyclogenesis in the west-central North Atlantic Ocean, 1981–84. Part I: Composite structure and mean behavior. *Mon. Wea. Rev.*, **114**, 1781–1794, doi:10.1175/1520-0493(1986)114<1781:ECITWC>2.0.CO;2.
- Sanders, F., 1988: Life history of mobile troughs in the upper westerlies. *Mon. Wea. Rev.*, **116**, 2629–2648, doi:10.1175/1520-0493(1988)116<2629:LHOMTI>2.0.CO;2.
- Sanders, F. and J. Gyakum, 1980: Synoptic-dynamic climatology of the “bomb”. *Mon. Wea. Rev.*, **108**, 1589–1606, doi:10.1175/1520-0493(1980)108<1589:SDCOT>2.0.CO;2.
- Sato, K., J. Inoue, A. Ya, J.-h. Kim, A. Maksh, V. Kustov, M. Maturilli, and K. Dethloff, 2018: Impact on predictability of tropical and mid-latitude cyclones by extra Arctic observations. *Sci. Rep.*, **8**, 12 104, doi:10.1038/s41598-018-30594-4.
- Schär, C. and H. Wernli, 1993: Structure and evolution of an isolated semi-geostrophic cyclone. *Quart. J. Roy. Meteor. Soc.*, **119**, 57–90, doi:10.1002/qj.49711950904.
- Schultz, D. M., 2022: The Presidents’ Day Snowstorm of 18–19 February 1979: Its history and significance at Monthly Weather Review. *Mon. Wea. Rev.*, **150**, 305–307, doi:10.1175/MWR-D-22-0012.1.

- Schwierz, C., M. Croci-Maspoli, and H. C. Davies, 2004: Perspicacious indicators of atmospheric blocking. *Geophys. Res. Lett.*, **31**, 1–4, doi:10.1029/2003gl019341.
- Shapiro, M. A., T. Hampel, and A. J. Krueger, 1987: The Arctic tropopause fold. *Mon. Wea. Rev.*, **115**, 444–454, doi:10.1175/1520-0493(1987)115<0444:TATF>2.0.CO;2.
- Shapiro, M. A., R. C. Schnell, F. P. Parungo, S. J. Oltmans, and B. A. Bodhaine, 1984: El Chichón volcanic debris in an Arctic tropopause fold. *Geophys. Res. Lett.*, **11**, 421–424, doi:10.1029/GL011i005p00421.
- Shimada, U., A. Wada, K. Yamazaki, and N. Kitabatake, 2014: Roles of an upper-level cold vortex and low-level baroclinicity in the development of polar lows over the Sea of Japan. *Tellus A: Dynamic Meteorology and Oceanography*, **66** (1), 24 694, doi:10.3402/tellusa.v66.24694.
- Shupe, M. D., 2011: Clouds at Arctic atmospheric observatories. Part II: Thermodynamic phase characteristics. *J. Appl. Meteor. Climatol.*, **50**, 645–661, doi:10.1175/2010JAMC2468.1.
- Simmonds, I. and I. Rudeva, 2012: The great Arctic cyclone of August 2012. *Geophys. Res. Lett.*, **39**, L230 709, doi:10.1029/2012GL054259.
- Simmonds, I. and I. Rudeva, 2014: A comparison of tracking methods for extreme cyclones in the Arctic basin. *Tellus*, **66A**, 25 252, doi:10.3402/tellusa.v66.25252.
- Sinclair, M. R., 1997: Objective identification of cyclones and their circulation intensity, and climatology. *Wea. Forecasting*, **12**, 595–612, doi:10.1175/1520-0434(1997)012<0595:OIOCAT>2.0.CO;2.
- Skamarock, W. C., M. G. Duda, S. Ha, and S. H. Park, 2018: Limited-area atmospheric modeling using an unstructured mesh. *Mon. Wea. Rev.*, **146**, 3445–3460, doi:10.1175/MWR-D-18-0155.1.
- Skamarock, W. C., J. B. Klemp, M. G. Duda, L. D. Fowler, S. H. Park, and T. D. Ringler, 2012: A multiscale nonhydrostatic atmospheric model using centroidal voronoi tessellations and C-Grid staggering. *Mon. Wea. Rev.*, **140**, 3090–3105, doi:10.1175/MWR-D-11-00215.1.
- Skamarock, W. C., J. B. Klemp, J. Dudhia, D. O. Gill, D. M. Barker, W. Wang, and J. G. Powers, 2005: A description of the advanced research WRF version 2. Tech. rep., NCAR, 88 pp.
- Stein, A. F., R. R. Draxler, G. D. Rolph, B. J. Stunder, M. D. Cohen, and F. Ngan, 2015: NOAA's HYSPLIT atmospheric transport and dispersion modeling system. *Bull. Amer. Meteor. Soc.*, **96**, 2059–2077, doi:10.1175/BAMS-D-14-00110.1.
- Stoelinga, M., 1996: A potential vorticity-based study of the role of diabatic heating and friction in a numerically simulated baroclinic cyclone. *Mon. Wea. Rev.*, **124**, 849–874, doi:10.1175/1520-0493(1996)124<0849:APVBSO>2.0.CO;2.

- Stoll, P. J., 2022: A global climatology of polar lows investigated for local differences and wind-shear environments. *Wea. Climate Dynam.*, **3** (2), 483–504, doi:10.5194/wcd-3-483-2022.
- Stoll, P. J., R. G. Graversen, G. Noer, and K. Hodges, 2018: An objective global climatology of polar lows based on reanalysis data. *Quart. J. Roy. Meteor. Soc.*, **144** (716), 2099–2117, doi:10.1002/qj.3309.
- Stoll, P. J., T. Spengler, A. Terpstra, and R. G. Graversen, 2021: Polar lows-moist-baroclinic cyclones developing in four different vertical wind shear environments. *Wea. Climate Dynam.*, **2** (1), 19–36, doi:10.5194/wcd-2-19-2021.
- Stoll, P. J., T. M. Valkonen, R. G. Graversen, and G. Noer, 2020: A well-observed polar low analysed with a regional and a global weather-prediction model. *Quart. J. Roy. Meteor. Soc.*, **146** (729), 1740–1767, doi:10.1002/qj.3764.
- Szapiro, N., 2019: Impacts of tropopause polar vortices on arctic sea ice loss. Phd thesis, University of Oklahoma, Norman, OK.
- Szapiro, N. and S. M. Cavallo, 2018: TPVTrack v1.0: A watershed segmentation and overlap correspondence method for tracking tropopause polar vortices. *Geosci. Model Dev.*, **11**, 5173–5187, doi:10.5194/gmd-11-5173-2018.
- Takayabu, I., 1991: “Coupling development”: An efficient mechanism for the development of extratropical cyclones. *J. Meteor. Soc. Japan*, **69**, 609–628.
- Takayabu, I., 1992: Supplement to “the coupling development”: Accelerative development of extratropical cyclones by using finite amplitude initial disturbances. *J. Meteor. Soc. Japan*, **70**, 837–841.
- Tao, W., J. Zhang, Y. Fu, and X. Zhang, 2017a: Driving roles of tropospheric and stratospheric thermal anomalies in intensification and persistence of the Arctic superstorm in 2012. *Geophys. Res. Lett.*, **44**, 10 017–10 025, doi:10.1002/2017GL074778.
- Tao, W., J. Zhang, and X. Zhang, 2017b: The role of stratosphere vortex downward intrusion in a long-lasting late-summer Arctic storm. *Quart. J. Roy. Meteor. Soc.*, **143**, 1953–1966, doi:10.1002/qj.3055.
- Thorpe, A. J., 1985: Diagnosis of balanced vortex structure using potential vorticity. *J. Atmos. Sci.*, **42**, 397–406, doi:10.1175/1520-0469(1985)042<0397:DOBVSU>2.0.CO;2.
- Thorpe, A. J., 1986: Synoptic scale disturbances with circular symmetry. *Mon. Wea. Rev.*, **114**, 1384–1389, doi:10.1175/1520-0493(1986)114<1384:SSDWCS>2.0.CO;2.
- Torn, R. D., 2014: The impact of targeted dropwindsonde observations on tropical cyclone intensity forecasts of four weak systems during PREDICT. *Mon. Wea. Rev.*, **142**, 2860–2878, doi:10.1175/MWR-D-13-00284.1.

- Torn, R. D. and G. J. Hakim, 2008: Ensemble-based sensitivity analysis. *Mon. Wea. Rev.*, **136**, 663–677, doi:10.1175/2007MWR2132.1.
- Turner, D. D., M. D. Shupe, and A. B. Zwink, 2018: Characteristic atmospheric radiative heating rate profiles in Arctic clouds as observed at Barrow, Alaska. *J. Appl. Meteor. Climatol.*, **57**, 953–968, doi:10.1175/JAMC-D-17-0252.1.
- Turner, J., E. A. Rasmussen, and A. M. Carleton, 2003: Introduction. *Polar Lows: Mesoscale Weather Systems in the Polar Regions*, E. A. Rasmussen and J. Turner, Eds., Cambridge University Press, 1–51, doi:10.1159/000184254.
- Uccellini, L. W., 1990: Processes contributing to the rapid development of extratropical cyclones. *Extratropical Cyclones: The Erik Palmén Memorial Volume*, C. Newton and E. O. Holopainen, Eds., Amer. Meteor. Soc., chap. 6, 81–105.
- Uccellini, L. W., D. Keyser, K. F. Brill, and C. H. Wash, 1985: The Presidents’ Day cyclone of 18–19 February 1979: Influence of upstream trough amplification and associated tropopause folding on rapid cyclogenesis. *Mon. Wea. Rev.*, **113**, 962–988, doi:10.1175/1520-0493(1985)113<0962:TPDCOF>2.0.CO;2.
- Uccellini, L. W., P. J. Kocin, R. A. Petersen, C. H. Wash, and K. F. Brill, 1984: The Presidents’ Day cyclone of 18–19 February 1979: Synoptic overview and analysis of the subtropical jet influencing the pre-cyclogenetic period. *Mon. Wea. Rev.*, **112**, 31–55, doi:10.1175/1520-0493(1984)112<0031:TPDCOF>2.0.CO;2.
- Valkonen, E., J. Cassano, and E. Cassano, 2021: Arctic cyclones and their interactions with the declining sea ice: A recent climatology. *J. Geophys. Res. Atmos.*, **126**, e2020JD034366, doi:10.1029/2020JD034366.
- Vessey, A. F., K. I. Hodges, L. C. Shaffrey, and J. J. Day, 2020: An inter-comparison of Arctic synoptic scale storms between four global reanalysis datasets. *Climate Dynam.*, **54** (5-6), 2777–2795, doi:10.1007/s00382-020-05142-4, URL <https://doi.org/10.1007/s00382-020-05142-4>.
- Vessey, A. F., K. I. Hodges, L. C. Shaffrey, and J. J. Day, 2022: The composite development and structure of intense synoptic-scale Arctic cyclones. *Wea. Climate Dynam.*, **3**, 1097–1112, doi:10.5194/wcd-3-1097-2022.
- Walsh, J. E., D. H. Bromwich, J. E. Overland, M. C. Serreze, and K. R. Wood, 2018: 100 years of progress in polar meteorology. *Meteor. Monogr.*, **59**, 21.1–21.36, doi:10.1175/AMSMONOGRAPHS-D-18-0003.1.
- Waugh, D. W., A. H. Sobel, and L. M. Polvani, 2017: What is the polar vortex and how does it influence weather? *Bull. Amer. Meteor. Soc.*, **98**, 37–44, doi:10.1175/BAMS-D-15-00212.1.

- Wernli, H. and C. Schwierz, 2006: Surface cyclones in the ERA-40 dataset (1958—2001). Part I: Novel identification method and global climatology. *J. Atmos. Sci.*, **63**, 2486–2507, doi:10.1175/JAS3766.1.
- Whitaker, J. S., L. W. Uccellini, and K. F. Brill, 1988: A model-based diagnostic study of the rapid development phase of the Presidents’ Day cyclone. *Mon. Wea. Rev.*, **116**, 2337–2365, doi:10.1175/1520-0493(1988)116<2337:AMBD SO>2.0.CO;2.
- Wirth, V., 1995: Diabatic heating in an axisymmetric cut-off cyclone and related stratosphere-troposphere exchange. *Quart. J. Roy. Meteor. Soc.*, **121**, 127–147, doi:10.1002/qj.49712152107.
- Wirth, V., 2001: Cyclone–anticyclone asymmetry concerning the height of the thermal and the dynamical tropopause. *J. Atmos. Sci.*, **58**, 26–37, doi:10.1175/1520-0469(2001)058<0026:CAACTH>2.0.CO;2.
- Wirth, V., M. Riemer, E. K. M. Chang, and O. Martius, 2018: Rossby wave packets on the midlatitude Waveguide—A review. *Mon. Wea. Rev.*, **146**, 1965–2001, doi:10.1175/MWR-D-16-0483.1.
- WMO, 1957: Meteorology—a three-dimensional science. *WMO Bull.*, 134–138.
- Yamagami, A., M. Matsueda, and H. L. Tanaka, 2017: Extreme Arctic cyclone in August 2016. *Atmos. Sci. Lett.*, **18**, 307–314, doi:10.1002/asl.757.
- Yang, C. C., C. C. Wu, K. H. Chou, and C. Y. Lee, 2008: Binary interaction between Typhoons Fengshen (2002) and Fungwong (2002) based on the potential vorticity diagnosis. *Mon. Wea. Rev.*, **136**, 4593–4611, doi:10.1175/2008MWR2496.1.
- Yoshida, A. and Y. Asuma, 2004: Structures and environment of explosively deepening extratropical cyclones in the northwest Pacific region. *Mon. Wea. Rev.*, **132**, 1121–1142, doi:10.1175/1520-0493(2004)132<1121:SAEOED>2.0.CO;2.
- Zhang, J., R. Lindsay, A. Schweiger, and M. Steele, 2013: The impact of an intense summer cyclone on 2012 Arctic sea ice retreat. *Geophys. Res. Lett.*, **40**, 720–726, doi:10.1002/grl.50190.
- Zhang, M., X. Y. Yang, and Y. Huang, 2022: Impacts of sudden stratospheric warming on extreme cold events in early 2021: An ensemble-based sensitivity analysis. *Geophys. Res. Lett.*, **49**, 1–10, doi:10.1029/2021GL096840.
- Zhang, X., J. E. Walsh, J. Zhang, U. S. Bhatt, and M. Ikeda, 2004: Climatology and interannual variability of Arctic cyclone activity: 1948–2002. *J. Climate*, **17**, 2300–2317, doi:10.1175/1520-0442(2004)017<2300:CAIVOA>2.0.CO;2.
- Zheng, M., et al., 2024: An assessment of dropsonde sampling strategies for atmospheric river reconnaissance. *Mon. Wea. Rev.*, **152**, 811–835, doi:10.1175/MWR-D-23-0111.1.

Zierl, B. and V. Wirth, 1997: The influence of radiation on tropopause behavior and stratosphere-troposphere exchange in an upper tropospheric anticyclone. *J. Geophys. Res.*, **102**, 883–894, doi:10.1029/97JD01667.