

The remarkably strong Arctic stratospheric polar vortex of Winter 2020: links to record-breaking arctic oscillation and ozone loss

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1 **The Remarkably Strong Arctic Stratospheric Polar**
2 **Vortex of Winter 2020: Links to Record-Breaking**
3 **Arctic Oscillation and Ozone Loss**

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15 **Key Points:**

- 16 • The Arctic stratospheric polar vortex during the 2019/2020 winter was the strongest
17 and most persistently cold in over 40 years
- 18 • Low tropospheric planetary wave driving and a wave-reflecting configuration of
19 the stratosphere supported the strong and cold polar vortex
- 20 • Seasonal records in the Arctic Oscillation and stratospheric ozone loss were related
21 to the strong polar vortex

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Abstract

The Northern Hemisphere (NH) polar winter stratosphere of 2019/2020 featured an exceptionally strong and cold stratospheric polar vortex. Wave activity from the troposphere during December-February was unusually low, which allowed the polar vortex to remain relatively undisturbed. Several transient wave pulses nonetheless served to help create a reflective configuration of the stratospheric circulation by disturbing the vortex in the upper stratosphere. Subsequently, multiple downward wave coupling events took place, which aided in dynamically cooling and strengthening the polar vortex. The persistent strength of the stratospheric polar vortex was accompanied by an unprecedentedly positive phase of the Arctic Oscillation in the troposphere during January-March, which was consistent with large portions of observed surface temperature and precipitation anomalies during the season. Similarly, conditions within the strong polar vortex were ripe for allowing substantial ozone loss: The undisturbed vortex was a strong transport barrier, and temperatures were low enough to form polar stratospheric clouds for over four months into late March. Total column ozone amounts in the NH polar cap decreased, and were the lowest ever observed in the February-April period. The unique confluence of conditions and multiple broken records makes the 2019/2020 winter and early spring a particularly extreme example of two-way coupling between the troposphere and stratosphere.

Plain Language Summary

Wintertime westerly winds in the polar stratosphere (from ~15-50km), known as the stratospheric polar vortex, were extraordinarily strong during the Northern Hemisphere winter of 2019/2020. The exceptional strength of the stratospheric polar vortex had consequences for winter and early spring weather near the surface, and for stratospheric ozone depletion. Typically atmospheric waves generated in the troposphere spread outward and upward into the stratosphere where they can disturb and weaken the polar vortex, but tropospheric wave activity was unusually weak during the 2019/2020 winter. In addition, an unusual configuration of the stratospheric polar vortex developed that reflected waves traveling upward from the troposphere back downward. These unique conditions allowed the vortex to remain strong and cold for several months. During January-March 2020, the strong stratospheric polar vortex was closely linked to a near-surface circulation pattern that resembles the positive phase of the so-called “Arctic Oscillation” (AO). This positive AO pattern was also of record strength, and influenced the regional distributions of temperatures and precipitation during the late winter and early spring. Cold and stable conditions within the polar vortex also allowed strong ozone depletion to take place, leading to lower ozone levels than ever before seen above the Arctic in spring.

1 Introduction

The Northern Hemisphere (NH) late winter and spring of 2020 featured a series of remarkable climate extremes. The tropospheric Arctic Oscillation – the dominant pattern of extratropical climate variability that describes the latitudinal shift of the eddy-driven jet stream (AO; Thompson & Wallace, 1998) – was effectively locked in a highly positive phase for several months. Stratospheric ozone in the polar cap fell to low levels never before observed in early NH spring. These phenomena were connected by the Arctic stratospheric polar vortex, which was unusually and persistently strong and cold during the season. This paper provides an overview of the 2019/2020 record breaking strong stratospheric polar vortex event and its connections to the extremes in the tropospheric AO and Arctic ozone.

During NH winter, the stratospheric and tropospheric circulations are closely connected. The principal circulation feature of the polar wintertime stratosphere is the stratospheric polar vortex (hereinafter, the polar vortex), which consists of a strong westerly circulation spanning from roughly 100 hPa to above 1 hPa (Waugh et al., 2017). Dur-

72 ing the winter polar night, the polar vortex strengthens and cools via radiative cooling.
 73 However, the strength of the polar vortex is also modulated by dynamical troposphere-
 74 stratosphere coupling via planetary scale waves generated in the troposphere from orog-
 75 raphy and sources of diabatic heating (e.g., Charney & Drazin, 1961; Matsuno, 1970).
 76 Waves from the troposphere can propagate vertically into the polar stratosphere, where
 77 they can break and disturb the polar vortex. Breaking waves deposit easterly momen-
 78 tum, which weakens the westerly zonal circulation represented by the polar vortex, and
 79 warms the polar stratosphere. Thus, the average strength of the polar vortex over a sea-
 80 son closely depends on the time-integrated wave driving of the stratosphere; for exam-
 81 ple, below average wave driving supports the development of a strong polar vortex, since
 82 uninterrupted radiative cooling allows the vortex to more closely approach the very cold
 83 conditions of radiative equilibrium.

84 Internal stratospheric processes can also influence polar vortex strength. Since wave
 85 propagation characteristics are determined by the basic state flow, the interplay between
 86 dynamic driving and radiative relaxation can alter the action of waves on the stratospheric
 87 circulation. For example, downward wave coupling events in which upward propagat-
 88 ing waves are reflected back from the stratosphere to the troposphere dynamically strengthen
 89 and cool the vortex by weakening or reversing the residual circulation (Shaw & Perlwitz,
 90 2014; Dunn-Sigouin & Shaw, 2015). These events have been shown to be preceded by
 91 transient pulses of upward wave activity that help develop reflective configurations of
 92 the polar stratospheric circulation (Harnik, 2009; Shaw et al., 2010; Shaw & Perlwitz,
 93 2013; Dunn-Sigouin & Shaw, 2018). Winters with more frequent downward wave cou-
 94 pling events generally correspond to winters with stronger polar vortices in the lower and
 95 middle stratosphere (Perlwitz & Harnik, 2003).

96 The interannual variability in the strength of the Arctic polar vortex is quite large.
 97 Sudden stratospheric warmings (SSWs) are relatively common in the NH, occurring in
 98 roughly 6 out of 10 years (Butler et al., 2017); these events involve an extreme mid-winter
 99 weakening of the polar vortex that is generally driven by enhanced wave driving. Since
 100 SSWs often lead to a nearly complete breakdown of the polar vortex, and the timescale
 101 of recovery from a weak stratospheric circulation can be long (Hitchcock & Shepherd,
 102 2013; Hitchcock et al., 2013), SSWs generally correspond to persistent weak polar vor-
 103 tex events. In contrast, persistent strong vortex events like that observed during the win-
 104 ter and spring of 2020 are quite rare in comparison to SSWs. Because of the relatively
 105 short timescales on which planetary wave driving acts, the polar vortex can rapidly shift
 106 from a strong state to a neutral or weak state (Limpasuvan et al., 2005; Lawrence & Man-
 107 ney, 2018). Maintaining a strong polar vortex for long periods of time thus requires unique
 108 conditions, such as weak upward wave activity and/or enhanced downward wave activ-
 109 ity.

110 The strength of the NH polar vortex is generally recognized as an important ele-
 111 ment for coupling between the stratosphere and troposphere on sub-seasonal to seasonal
 112 timescales during winter and spring (e.g., Kidston et al., 2015; Butler et al., 2019). A
 113 main expression of two-way stratosphere-troposphere dynamical coupling during NH win-
 114 ter is the close statistical relationship between the strength of the stratospheric polar vor-
 115 tex and the phase of the tropospheric AO (e.g., Baldwin & Dunkerton, 2001; Kidston
 116 et al., 2015). These relationships are commonly expressed using metrics that describe
 117 phases of the “Northern Annular Mode” (NAM), a pattern that characterizes meridional
 118 shifts of mass into or out of the polar cap throughout the atmospheric column (note that
 119 the NAM and AO are often used interchangeably; Thompson & Wallace, 2000; Baldwin,
 120 2001). Anomalously strong or weak polar vortex states correspond to positive or neg-
 121 ative phases of the stratospheric NAM, respectively, and these tend to be followed in the
 122 troposphere by positive or negative AO events, which may last for weeks to months and
 123 alter patterns of surface temperatures and precipitation (Baldwin & Dunkerton, 2001;
 124 Polvani & Kushner, 2002; Limpasuvan et al., 2005; Dunn-Sigouin & Shaw, 2015; Kid-

125 ston et al., 2015; Tripathi, Charlton-Perez, et al., 2015; Orsolini et al., 2018; Domeisen,
126 2019; King et al., 2019). Downward wave coupling events can not only strengthen the
127 polar vortex, but also directly induce tropospheric circulation patterns consistent with
128 a positive AO on short timescales (Shaw & Perlwitz, 2013; Dunn-Sigouin & Shaw, 2015).
129 However, phases of the tropospheric AO/NAM do not always consistently follow the strength
130 of the polar vortex. Factors that seem to determine whether a given vortex event will
131 influence the troposphere include the persistence and magnitude of stratospheric anom-
132 alies, the depth to which anomalies penetrate into the lower stratosphere, and the tropo-
133 spheric state at the time of the stratospheric event (Kodera et al., 2016; Karpechko et
134 al., 2017; Charlton-Perez et al., 2018; Domeisen, 2019; White et al., 2019; Rao et al., 2020).

135 The conditions that determine the potential for chemical ozone destruction in the
136 NH stratosphere also tie in to polar vortex strength, albeit in subtle ways that are highly
137 sensitive to meteorology (WMO, 2014, 2018). Chlorine and bromine trace gases, primar-
138 ily from anthropogenic sources, are converted from reservoir (non-ozone depleting) forms
139 to reactive (ozone-depleting) forms on the surfaces of polar stratospheric clouds (PSCs;
140 e.g., Solomon, 1999), which require very low temperatures (~ 195 K) to form in the lower
141 stratosphere. Activation of chlorine/bromine also generally requires persistent confine-
142 ment with cold air inside the polar vortex so that mixing with low latitude air cannot
143 dilute the “activated air” (Schoeberl & Hartmann, 1991; Schoeberl et al., 1992). The chem-
144 ical reactions that destroy ozone further require sunlight exposure, such that chemical
145 ozone loss tends to dominate when sunlight returns to the polar regions in early spring,
146 a time when, climatologically, the Arctic vortex is often very weak or broken down al-
147 together (Black et al., 2006; Lawrence et al., 2018). The aforementioned conditions for
148 ozone destruction are typically only present when the polar vortex is strong, cold, and
149 stable, but the interannual variability in the Arctic polar vortex is so large that individ-
150 ual seasons can have individual conditions present without the others: For example, the
151 polar vortex in 2015/2016 was persistently strong and cold for much of the season, but
152 a dynamically driven early final warming occurred in the beginning of March, which cut
153 short the chemical ozone loss, and broke down the vortex (Manney & Lawrence, 2016),
154 preventing an extreme ozone deficit. Downward wave coupling events in the stratosphere
155 encourage chemical ozone loss through dynamically cooling and strengthening the po-
156 lar vortex; they also reduce the downward resupply of ozone through their ability to weaken
157 and/or reverse the residual circulation (Shaw & Perlwitz, 2014; Lubis et al., 2017).

158 In this paper we will show that the 2019/2020 record breaking strong vortex devel-
159 oped in the wake of a combination of low wave driving from the troposphere and mul-
160 tiple downward wave coupling events that occurred following formation of a reflective
161 configuration in the upper stratospheric circulation. The record-breaking strength of the
162 vortex was accompanied by a record-breaking positive phase of the tropospheric AO that
163 lasted several months and was related to large fractions of NH seasonal surface temper-
164 atures and precipitation anomalies. We will further illustrate that the strong and sta-
165 ble vortex also provided conditions that were ideal for chemical ozone loss to take place,
166 resulting in the lowest Arctic ozone amounts on record during late winter and early spring.
167 That the record-breaking AO and low ozone events took place individually is notable,
168 but that they both occurred during the same season makes the 2019/2020 Arctic win-
169 ter particularly extraordinary.

170 The rest of the paper is organized as follows: Section 2 outlines the datasets and
171 methods we use. Section 3 is broken into subsections that focus on describing the record
172 strength of the vortex (Section 3.1); the coupled troposphere-stratosphere evolution (Sec-
173 tion 3.2); the influence of two-way wave coupling on the vortex (Section 3.3); and the
174 vortex conditions that were conducive for ozone loss (Section 3.4). In Section 4, we briefly
175 discuss our results in the context of previous winters, and provide some research ques-
176 tions that are motivated by this record-breaking winter and early spring. Finally, in Sec-
177 tion 5 we summarize our results.

2 Data and Methods

We combine data from multiple sources to analyze the conditions during the 2019/2020 Arctic winter, and to provide historical context from previous winters. Meteorological variables such as temperatures, winds, and geopotential height are from the National Aeronautics and Space Administration (NASA) Modern-Era Retrospective analysis for Research and Applications version 2 (MERRA-2; Gelaro et al., 2017). We specifically use daily mean fields from the pressure (“M2I3NPASM”; GMAO, 2020b) and model (“M2I3NVASM” GMAO, 2020a) level collections. For historical context of stratospheric zonal mean zonal winds from previous winters, we also utilize daily mean pressure level data from the Japanese Meteorological Agency’s 55-year reanalysis (JRA-55; Kobayashi et al., 2015) for winter seasons from 1958/1959 to 1978/1979. Ozone data and statistics are compiled from multiple satellite instruments, but are primarily from the Ozone Mapping and Profiling Suite (OMPS) from data made available via the NASA OzoneWatch resource (see, e.g., <https://ozonewatch.gsfc.nasa.gov/data/> and <https://ozonewatch.gsfc.nasa.gov/meteorology/figures/ozone/>); missing column ozone values in polar night are filled using MERRA-2 data. Daily values for the Arctic Oscillation index are provided by the National Centers for Environmental Prediction (NCEP) Climate Prediction Center (CPC) at https://www.cpc.ncep.noaa.gov/products/precip/CWlink/daily_ao_index/ao.shtml; we refer to these data as the AO_{CPC} .

We use diagnostics based on the Transformed Eulerian Mean (TEM) framework (Andrews et al., 1987), including Eliassen-Palm fluxes and residual velocities to describe the wave driving conditions and evolution of the stratospheric circulation during the 2019/2020 winter season. We calculate these diagnostics based on the primitive equation formulation (see, e.g., Martineau et al., 2018) using MERRA-2 pressure level fields. We also use diagnostics of polar processing, which describe the development and maintenance of conditions that support chemical ozone loss; we compute these as described in Lawrence et al. (2018) using daily mean MERRA-2 data. Briefly, we use isentropic potential vorticity (PV) to determine the size of the polar vortex and the magnitude of PV gradients at the vortex edge, characteristics that assess the polar vortex as a transport barrier. We also use temperatures to determine whether conditions support the development of PSCs, and the size of regions able to form PSCs. We specifically express the size of regions cold enough to form nitric acid trihydrate (NAT) PSCs as the volume of cold air divided by the volume of the vortex (V_{NAT}/V_{vort}), where the volumes span only the lower stratosphere (see Lawrence et al., 2018, for details).

Unless otherwise noted, we calculate anomalies with respect to climatologies using the full records available, but excluding 2020. Similarly, we use cosine-latitude weighted averages to calculate quantities representative of a range of latitudes. Note that the NAM and AO refer to identical phenomena (Baldwin, 2001; Baldwin & Dunkerton, 2001), but herein we use the NAM to refer to the vertically resolved profile of mass fluctuations in the NH extratropical circulation, and the AO to refer to the near-surface pattern. We calculate the vertically resolved NAM index using standardized 65-90°N geopotential height anomalies as motivated by Cohen et al. (2002) and Baldwin and Thompson (2009), multiplied by -1 for consistent phasing with the AO.

3 Results

3.1 Strength of the 2019/2020 Polar Vortex in Context

In the middle stratosphere, zonal mean zonal winds were above average between 55-75°N for the majority of the extended winter season, but became particularly strong around mid-January (Figure 1a). Beginning in January, polar vortex winds were regularly more than 20 m/s higher than those in the climatology. In February, the wind anomalies exceeded two standard deviations of the November-April climatology for over a full

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month and reached record maxima during a period of time in the seasonal cycle when winds in this altitude and latitude region generally decrease.

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The temporal evolution of zonal wind anomalies at 60°N as a function of pressure reveals that the vortex was generally stronger than normal in the stratosphere between 100 and 1 hPa from November to April (Fig 1b). The only exception is a short-lived vortex disturbance from mid-November to early December, as evidenced by negative wind anomalies between about 30 and 1 hPa at this time. Winds in the troposphere became anomalously positive for a brief period in early December, while more consistent positive anomalies that often reached more than 10 m/s above normal became established in January.

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Also notable is the zonal wind evolution in the upper stratosphere and lower mesosphere (USLM; approximately pressures lower than 1 hPa). Following the short lived stratospheric vortex disturbance in mid-November, winds in the USLM accelerated and briefly became very strong, reaching record high values and exceeding 2 standard deviations for a short time in mid-December. However, beginning in January, there is a clear contrast between winds in the USLM and the stratosphere; those in the USLM were generally weaker than normal, while those in the stratosphere proper were generally stronger than normal, and reached record strength for periods in February and March.

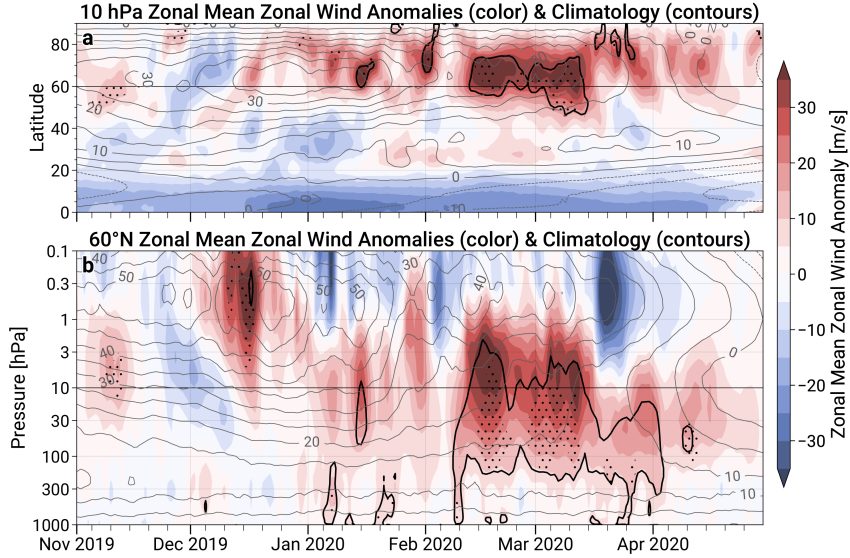


Figure 1. Time series of zonal mean zonal wind anomalies as a function of latitude at 10 hPa (a), and at 60°N as a function of pressure (b). The grey line contours represent the climatology; the black lines enclose the times when anomalies exceeded +2 standard deviations of the November-April daily climatology; and stippling indicates when the zonal wind values were maxima in the MERRA-2 record.

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The stratospheric circulation was clearly stronger than normal for almost the entirety of the extended December-March (DJFM) winter season. A comparison of zonal mean zonal winds across other winter seasons reveals that the polar vortex in 2020 was the strongest on record at 10 and 100 hPa for seasons back to 1979/1980 (Figure 2). This era is typically considered to be the “satellite-era”; when also including prior years back to 1958/1959 for which reanalysis data are more uncertain because of the relative lack of observations to constrain the reanalysis (see discussion in Hitchcock, 2019), the 2020

253 zonal winds at 10 hPa rank third across all available years, only exceeded by 1966/1967
 254 and 1975/1976. At 100 hPa, the 2019/2020 zonal winds are the largest on record even
 255 when taking into account these earlier years. We note that in the post-1980 era, the dif-
 256 ferences in the seasonal zonal winds between MERRA-2 and JRA-55 are very small; the
 257 absolute maximum differences in the DJFM means are 0.6 m/s and 1.0 m/s at 10 and
 258 100 hPa, respectively, indicating that these results are robust between these two reanal-
 259 ysis data sets. These results also demonstrate that the rankings for seasonal strength of
 260 the polar vortex in the middle stratosphere do not always correspond to those in the low-
 261 ermost stratosphere. For example, the years that follow 2019/2020 in ranking for sea-
 262 sonally strong polar vortices at 10 hPa such as 1995/1996, 1996/1997, and 2010/2011
 263 have values at 100 hPa that are exceeded by other years such as 1989/1990 and 1992/1993.

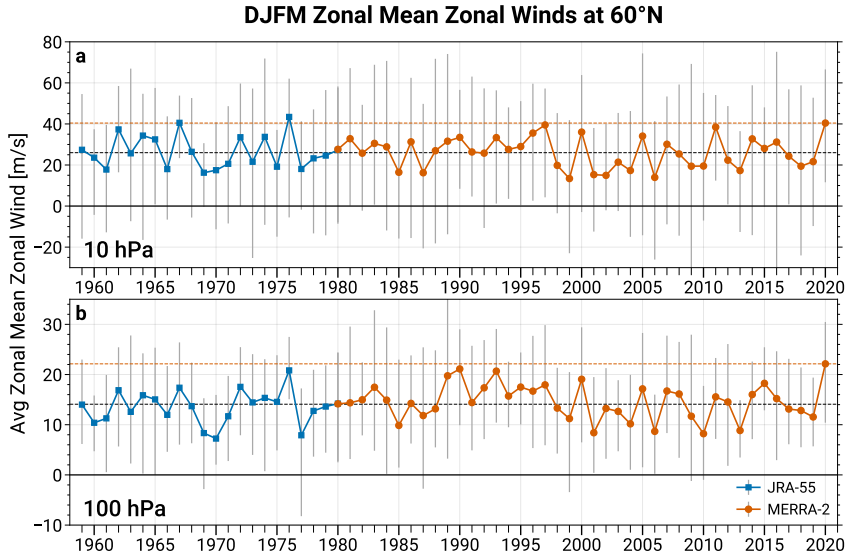


Figure 2. Yearly time series of the December-March averaged zonal mean zonal winds at $60^\circ N$, at 10 (a) and 100 (b) hPa. The blue lines and squares represent values determined from the JRA-55 reanalysis for 1959 through 1979; the orange lines and circles represent the values determined from MERRA-2. The grey whiskers in each panel represent the range of the daily mean zonal wind values during each season.

264 **3.2 An Extreme Event of the Coupled Troposphere-Stratosphere An-**
 265 **ular Mode**

266 The 2020 strong vortex event that developed in January and lasted through March
 267 was vertically coherent throughout the depth of the stratosphere. Moreover, the posi-
 268 tive zonal wind anomalies in the troposphere during this time indicate that the zonal pat-
 269 tern also extended into the troposphere (Figure 1). Figures 3a and b show the coherent
 270 evolution of stratospheric and tropospheric circulation anomalies characterized by indices
 271 of the NAM and AO, which clearly illustrate a positive NAM/AO state between 1000
 272 and 1 hPa for almost the entire three months of January-March (JFM).

273 We use two diagnostics to illustrate how unusual this winter was with respect to
 274 the coupled stratosphere-troposphere NAM behaviour. First, we assess the influence of
 275 wave driving on the stratospheric polar vortex. Newman et al. (2001) showed that early
 276 spring polar stratospheric temperatures are highly correlated with time integrated eddy

277 heat fluxes, revealing that interannual variability in spring polar stratospheric temper-
 278 atures is tied to the integrated amount of wave driving supplied by the troposphere and
 279 entering the stratosphere. Similarly, Polvani and Waugh (2004) showed a robust anti-
 280 correlation between time integrated eddy heat fluxes and the stratospheric NAM, fur-
 281 ther indicating a control on the vortex strength by wave driving. Figure 3c supplements
 282 these relationships by displaying a scatterplot of the 100 hPa 40-80°N vertical compo-
 283 nent of the Eliassen-Palm (EP) flux (F_z ; a diagnostic of vertical wave propagation) av-
 284 eraged over DJF versus the 50 hPa NAM averaged over JFM, which confirms a very close
 285 relationship ($r = -0.8$). Moreover, Figure 3c clearly illustrates that the 2020 winter sea-
 286 son represents a new extreme, with both the lowest DJF upward wave activity at 100
 287 hPa and the strongest 50 hPa NAM event in the MERRA-2 record.

288 Second, we put the 2020 coherent stratospheric and tropospheric NAM/AO behav-
 289 ior into context with previous years. Prior studies have shown that there is a significant
 290 statistical relationship between the strength of the stratospheric polar vortex (stratospheric
 291 NAM) and the AO on seasonal timescales (e.g., Thompson & Wallace, 1998). Figure 3d
 292 demonstrates this relationship as a scatterplot of JFM values of the 50 hPa NAM ver-
 293 sus polar cap sea level pressure (SLP). The correlation is approximately -0.68 , and is sta-
 294 tistically significant at the 99% level following a bootstrap test of 50000 resamples. The
 295 JFM season of 2020 particularly stands out as the most extreme year in the MERRA-
 296 2 record, involving extremes in both the stratospheric NAM and negative sea level pres-
 297 sure anomalies. While this result does not imply a clear direction of influence or causal-
 298 ity, it is obvious from Figure 3a that the stratospheric anomalies were persistent, of large
 299 magnitude, and reached into the lower stratosphere. Similarly, a positive AO developed
 300 slightly before or simultaneous with the stratospheric anomalies in late December and
 301 early January, meaning that the tropospheric anomalies either developed in concert with
 302 the stratosphere, or was in a favorable state for coupling with a positive stratospheric
 303 NAM.

304 While we have shown that the 2020 JFM NAM index was consistent with extremely
 305 low upward wave activity at 100 hPa (Fig 3c), the 100 hPa level is generally represen-
 306 tative of the lower stratosphere, and thus upward wave activity at this level is not nec-
 307 essarily indicative of wave activity from the troposphere (e.g., see discussion in de la Cámara
 308 et al., 2017). Figure 4 shows the yearly DJF mean F_z at 300 hPa in the upper tropo-
 309 sphere versus 100 hPa as a scatterplot. These are positively correlated, but only mod-
 310 estly so ($r = 0.46$), indicating that the amount of wave activity in the upper troposphere
 311 is not a perfect predictor of that for the lower stratosphere on seasonal timescales. Nonethe-
 312 less, 2019/2020 stands out among the other years as being the most coherent extreme
 313 minimum in DJF F_z at both 100 and 300 hPa. This result ties back to the NAM and
 314 SLP relationships illustrated in Figure 3, indicating that on average low upward wave
 315 driving of the stratosphere by the troposphere likely played a role in the development
 316 of the strong polar vortex in JFM (Fig 3c), and subsequently the negative polar cap SLP
 317 anomalies (Fig 3d).

318 At the surface, extratropical SLP anomalies were consistent with the long-lived posi-
 319 tive AO and strong stratospheric polar vortex (Fig 3a,b,d). Figure 5a shows that the
 320 SLP anomalies throughout JFM were primarily characterized by an annular pattern of
 321 anomalously low pressure in the polar cap, surrounded by a ring of anomalously high
 322 pressure in mid-latitudes, which closely resembles the canonical AO pattern. Figure 5b
 323 illustrates the 2020 JFM mean AO_{CPC} index was the highest on record since 1950 with
 324 a value of ~ 2.7 . Moreover, the persistence of this positive AO event was unprecedented;
 325 the minimum and maximum daily AO_{CPC} index values during JFM 2020 were both the
 326 highest on record, and values were consecutively above 1 for 56 days, greater than any
 327 previous year shown (Fig 5c). The JFM seasons of 1988/1989 and 1989/1990 also fea-
 328 tured large and persistently positive AO events; both of these years also featured polar

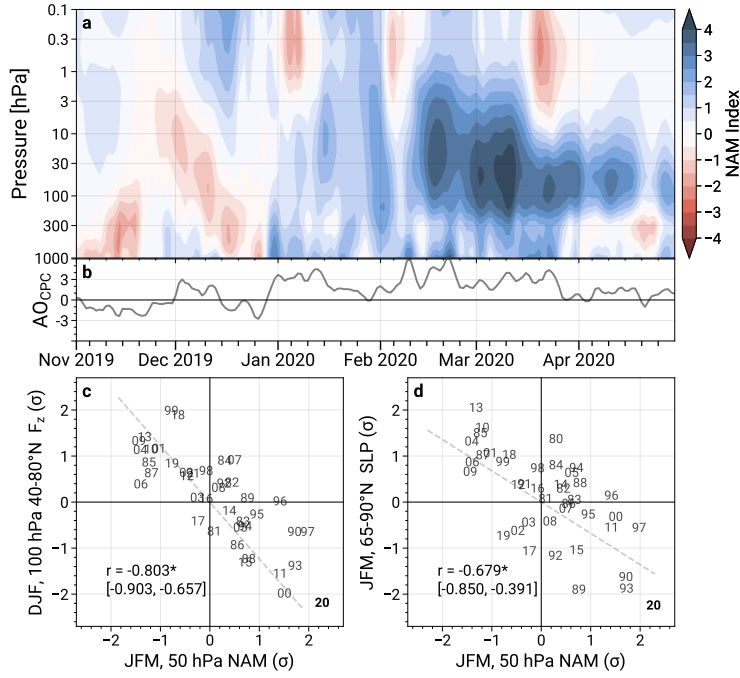


Figure 3. Time series of the Northern Annular Mode (a) and CPC Arctic Oscillation (b) indices from November 2019 through April 2020. Also shown are scatterplots of December-February (DJF) 100 hPa 40-80°N averaged vertical component of the Eliassen-Palm Flux (F_z) versus the JFM 50 hPa NAM index (c), and the JFM 50 hPa NAM index versus 65-90°N sea level pressure (d). All quantities in the scatter plots are standardized with respect to the yearly seasons. Correlations are indicated in the bottom left of panels c and d above 99% bootstrap confidence intervals from 50000 resamples.

329 vortices of above average seasonal strength in the lower stratosphere (particularly 1989/1990;
 330 see Figures 2b and 3d).

331 The extreme positive AO event that occurred during JFM 2020 explains a substan-
 332 tial fraction of the observed surface temperature and precipitation anomalies, including
 333 record warmth that occurred in Eurasia. Figure 6 compares the observed seasonal pat-
 334 terns of surface temperature and precipitation anomalies with those that are congruent
 335 with the AO, determined from multiplying the 2020 JFM AO_{CPC} value with the regres-
 336 sion map of these quantities onto the JFM AO_{CPC} historical time series. Surface tem-
 337 peratures were primarily characterized by very anomalous warmth in Eurasia, and cold
 338 in Canada, Greenland, and Alaska (Fig 6a). The Eurasian warmth (from 0-135°E, 45-
 339 75°N) was unprecedented in the MERRA-2 record back to 1980 (not shown). Precip-
 340 itation was largely above normal in bands along Northern Europe, central Siberia, and
 341 southern Eurasia (Fig 6d). The patterns congruent with the AO are generally consist-
 342 ent with that observed, but typically of lesser amplitude (e.g., the underestimation of
 343 temperatures over Eurasia; Fig 6b,e). Zonal means of the observed and AO-congruent
 344 anomalies (Fig 6c,f) highlight rough estimates of the fractions of patterns attributable
 345 to the AO. Between 40 and 70°N, the JFM AO explains about 2/3 of the amplitude of
 346 temperature anomalies, with a residual of about 0.5 K. The AO explains virtually all of
 347 the zonal mean precipitation anomalies between roughly 55-70°N, but overestimates the
 348 dry band along approximately 40°N. We note these quantities are not detrended, and

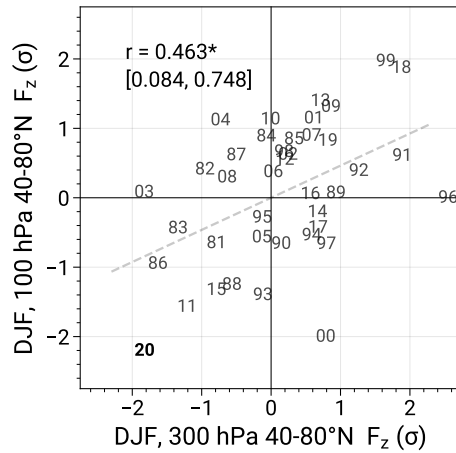


Figure 4. Scatterplot of the December-February (DJF) mean of the 40-80°N averaged vertical component of the EP-flux (F_z) at 300 hPa versus 100 hPa. The values shown are standardized with respect to the yearly seasons. The year labels are for the January of each season. The correlation is indicated in the top left above 99% bootstrap confidence intervals from 50000 resamples.

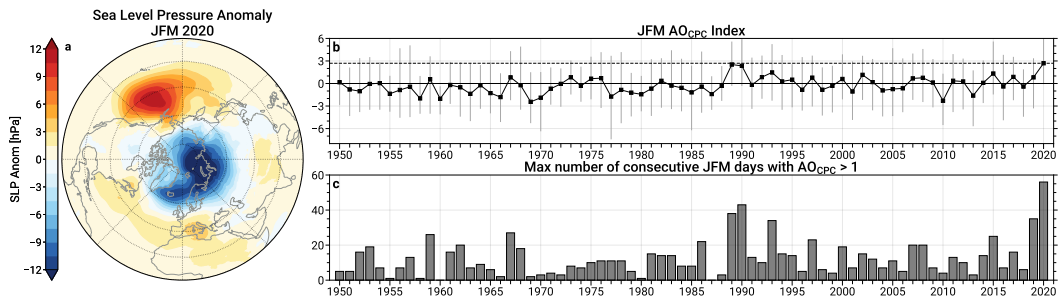


Figure 5. Map of Northern Hemisphere sea level pressure anomalies averaged over January-March (JFM) 2020 (a), yearly time series of the JFM mean CPC AO index (b), and yearly time series of the max number of consecutive JFM days in which the CPC AO index exceeded 1 (c). The whiskers in panel b represent the range of the AO values during the respective JFM seasons; the black dashed horizontal line is plotted at the mean value for 2020.

349 thus some of the observed patterns (such as the Eurasian warmth) may also be attributable
 350 to climate change warming.

351 **3.3 Wave Driving and Reflection: Dynamic Control of Polar Vortex Strength**

352 The previous subsection clearly illustrated the unusual conditions of the coupled
 353 stratosphere-troposphere system over the 2019/2020 winter season. Now we will describe
 354 in more detail the processes that led to the development of such a strong polar vortex
 355 by focusing more closely on the wave driving conditions.

356 The occurrence of the extremely strong stratospheric polar vortex of 2020 can be
 357 partly understood though a closer examination of the evolution of tropospheric wave driv-
 358 ing throughout the season (Figure 7). In general, waves in the troposphere that linearly
 359 interfere in a constructive/destructive way with the climatological stationary wave pat-

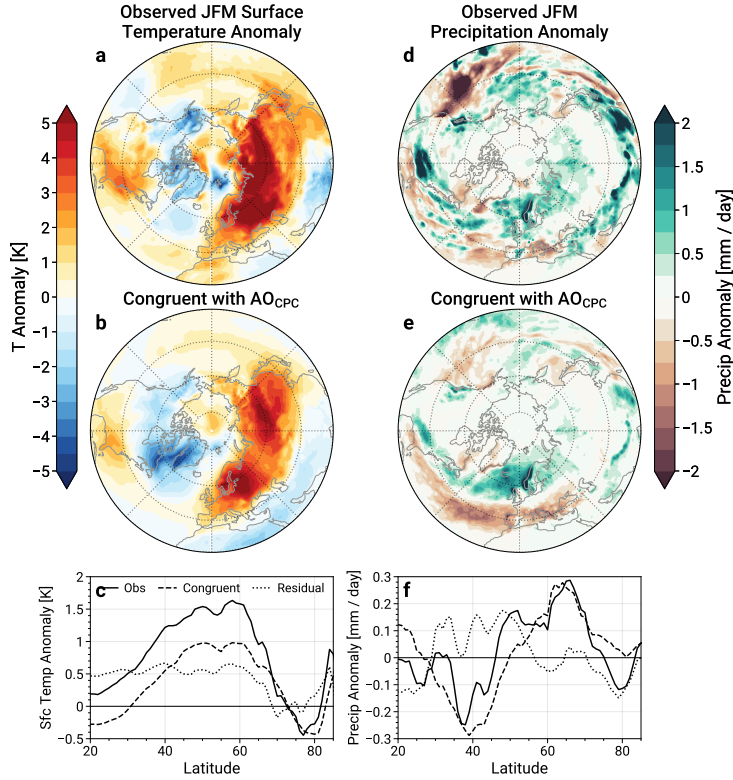


Figure 6. Maps of the observed January-March (JFM) 2020 anomalies in surface temperatures and precipitation (a,d), and the anomalies congruent with the JFM AO_{CPC} (b,e). The last row shows the zonal means of the observed anomalies, the AO reconstruction, and the residuals (c, f).

tern result in amplified/dampened wave driving of the polar vortex (see, e.g., Garfinkel et al., 2010; Kolstad & Charlton-Perez, 2011; Smith & Kushner, 2012). Figure 7a-e shows maps of the monthly 300 hPa geopotential height anomalies during the 2019/2020 season superposed with the climatological stationary wave patterns. November 2019 (Fig 7a) featured enhanced ridging over the Gulf of Alaska and the Ural mountains region. The patterns of 300 hPa geopotential height anomalies were generally constructive with the climatological stationary waves, which indicates enhanced wave driving occurring during this time. This is consistent with the positive anomalies in 40-80°N F_z (Fig 7f) in the troposphere and stratosphere from mid to late November, which were associated with a short duration vortex weakening event (see, e.g., Figures 1 and 3). The December geopotential height anomalies (Fig 7b) show less coherent interference patterns, which is consistent with the alternating periods of positive and negative F_z anomalies within the troposphere. In contrast, January 2020 featured geopotential height anomaly patterns in a configuration that destructively interfered with the climatological stationary waves, particularly over North America and the Pacific ocean. January also had persistently low values of F_z in both the troposphere and stratosphere, indicating a prolonged period of low upward wave activity in the stratosphere. Geopotential height anomalies during February and March 2020 (Fig 7d,e) primarily show the canonical development of the positive NAM/AO state, with negative anomalies in the polar cap, and positive anomalies in the midlatitudes, similar to the SLP pattern shown in Figure 5. We showed above that upward wave activity averaged over DJF was anomalously low in the troposphere and stratosphere (Figures 3 and 4). However, there are several periods through-

382 out the extended 2019/2020 season when F_z was anomalously high, particularly in the
 383 stratosphere, such as in mid-to-late November, mid-December to early January, late Janu-
 384 ary/early February, and mid-March (Fig 7f).

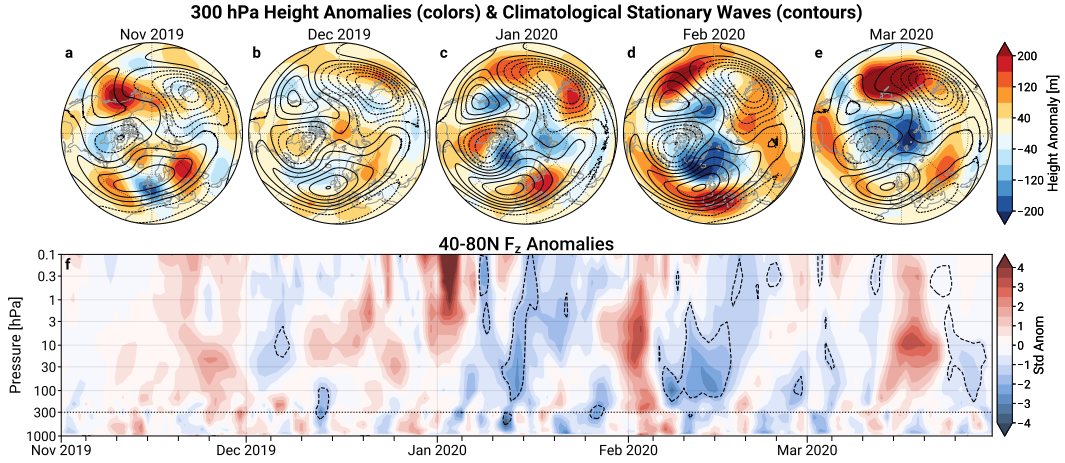


Figure 7. Maps of monthly 300 hPa geopotential height anomalies (color fill) and climatological eddy heights representing the climatological stationary waves for November 2019 – March 2020 (a - e). The bottom row (f) shows the daily time series of standardized anomalies in the 40 – 80°N average upward component of the Eliassen-Palm flux (F_z ; values are standardized using only October – March anomalies). Contours for eddy heights in the maps of a - e are plotted every 40m for values between -200 and 200m. Dashed contours in panel f show the times when the 40 – 80°N average meridional heat flux was negative.

385 Somewhat paradoxically, the transient positive F_z anomalies indicative of enhanced
 386 wave activity in the stratosphere likely played a role in promoting the robust polar vortex
 387 during the 2019/2020 season. The dashed contours in Figure 7f indicate when the
 388 40-80°N averaged meridional eddy heat flux ($\overline{v'T'}$) was negative. The vertical compo-
 389 nent of the EP-Flux, F_z , involves a term proportional to the eddy heat flux and tends
 390 to be dominated by it (Andrews et al., 1987); therefore, the prolonged periods of nega-
 391 tive stratospheric heat fluxes in January, February, and March were generally periods
 392 of time when wave propagation was downward as opposed to upward, indicative of wave
 393 reflection. The low seasonal F_z values shown in Figures 3c and 4, particularly at 100 hPa
 394 are thus partly a manifestation of averaging over enhanced *downward* wave activity, not
 395 just less *upward* wave activity.

396 It is well known that wave-mean flow interactions with planetary scale waves drive
 397 wintertime polar stratospheric temperatures away from radiative equilibrium; the depo-
 398 sition of easterly momentum by upward propagating planetary waves establishes a merid-
 399 ional residual circulation, which drives a polar downwelling that adiabatically warms the
 400 polar stratosphere (e.g., Andrews et al., 1987). However, total negative heat flux events
 401 which involve downward wave propagation, can have an episodic effect on the residual
 402 circulation by causing it to reverse with upward motion in the polar cap, leading to tran-
 403 sient adiabatic cooling of the polar stratosphere and strengthening of the polar vortex
 404 (Shaw & Perlwitz, 2013, 2014). These kinds of downward wave coupling events prefer-
 405 entially occur when the configuration of stratospheric winds support wave reflection, par-
 406 ticularly for zonal wavenumber-1 waves (Perlwitz & Harnik, 2003; Harnik, 2009; Shaw
 407 et al., 2010; Shaw & Perlwitz, 2013).

408 The zonal wind pattern in mid- and late winter 2020 evolved into such a reflective
 409 configuration. Figures 8a-e show monthly mean zonal mean zonal winds and EP-Flux
 410 vectors. Zonal winds in November and December (Fig 8a,b) primarily featured a single
 411 broad stratospheric jet with positive zonal wind shear over much of the extratropics. The
 412 average EP-Flux vectors during this time indicate wave propagation within the regions
 413 of strong westerlies through the stratosphere, with equatorward propagation inhibited
 414 by the regions of easterlies in the tropical stratosphere. Beginning in January and per-
 415 sisting through March (Fig 8c,d,e), a “split” jet structure emerged involving a high lat-
 416 itude jet maximum (around 60-70°N) in the lower to upper stratosphere, and a low lat-
 417 itude subtropical jet maximum (around 30-40°N) in the USLM. This configuration of
 418 the polar vortex features strong curvature of the zonal winds, a zonal wind minima
 419 in the lower and middle stratosphere that extends from low to mid-latitudes, and negative
 420 zonal wind shear at latitudes around 60°N in the middle to upper stratosphere (see also
 421 Fig 1b). This configuration has been shown to be highly reflective for stationary wavenumber-
 422 1 waves because the zonal wind minima in the low-mid latitude lower and middle strato-
 423 sphere act to meridionally confine waves, and the strong negative zonal wind shear acts
 424 as a vertical “cap” beyond which wave propagation is impaired (Perlwitz & Harnik, 2003;
 425 Harnik, 2009; Shaw et al., 2010). Since reflection events are relatively transient, the monthly-
 426 average EP-Flux vectors generally do not show signs of wave reflection (downward point-
 427 ing arrows) over the months of January – March; however, they do demonstrate the verti-
 428 cal cap in the high-latitude regions of negative zonal wind shear where wave propaga-
 429 tion is inhibited (particularly in Fig 8c,d), despite the winds being westerly.

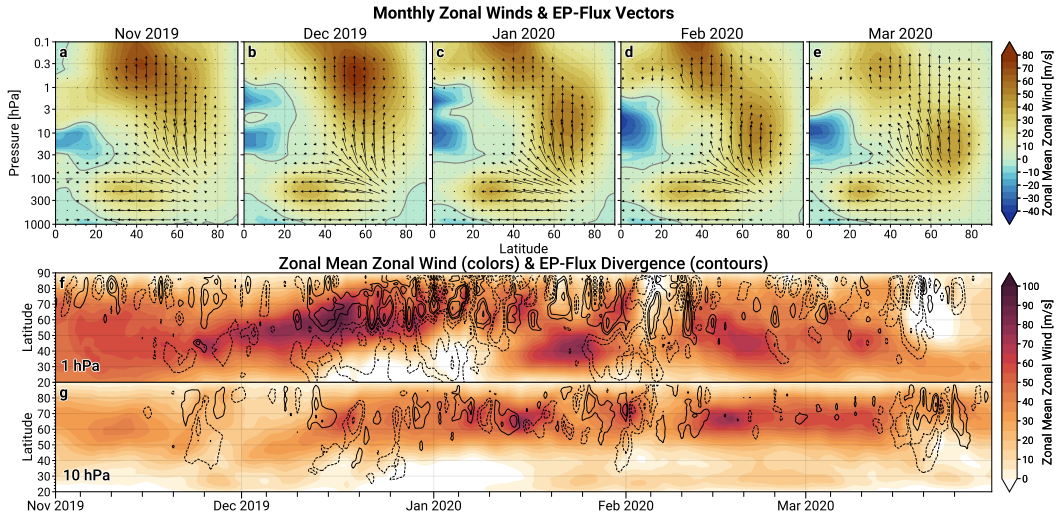


Figure 8. Latitude-pressure cross-sections of monthly zonal mean zonal winds and EP-flux vectors for November 2019 – March 2020 (a – e). The two bottom rows show latitude time series of zonal mean zonal winds at 1 (f) and 10 (g) hPa with contours of the acceleration by the EP-flux divergence overlaid. Only relatively extreme values of EP-flux divergence are plotted, for contours of $\pm[8, 16, 32, 64]$ m/s/day (contours for 0 m/s/day are excluded).

430 This split-jet polar vortex structure initially developed following a transient dis-
 431 turbance in early January that primarily affected the vortex within the USLM (see Fig 7f).
 432 Figure 8f,g show latitude/time series of zonal winds and acceleration by EP-Flux diver-
 433 gence from November through March at 10 and 1 hPa. While the jet maximum at 1 hPa
 434 began the season at relatively low latitudes around 40°N, it shifted poleward under wave
 435 driving before being nearly eroded away in early January. Due to the decreases in den-

436 sity with altitude, waves that reach the upper stratosphere tend to grow to large ampli-
 437 tudes and break there, resulting in warming of the polar upper stratosphere, and a pole-
 438 ward movement of the vortex edge like that shown here (Dunkerton & Delisi, 1986; Dunker-
 439 ton, 2000; Scott et al., 2004). However, radiative time scales are short at these altitudes
 440 (e.g., Newman & Rosenfield, 1997), meaning that fast cooling under radiative relaxation
 441 can allow the rapid re-establishment of the upper stratospheric jet maximum at lower
 442 latitudes (e.g., Dunkerton & Delisi, 1985; Dunkerton, 2000). This process is consistent
 443 with the zonal wind evolution at 1 hPa (and higher altitudes; not shown) in January,
 444 and it repeated in February. The polar vortex jet at 10 hPa remained comparatively undis-
 445 turbed during these times (Fig 8g) due to the transient nature of the upward wave pulses,
 446 meaning negative wind shear developed between the middle and upper stratosphere around
 447 60-70°N (associated with the upper-level negative wind anomalies in Fig 1b). The neg-
 448 ative heat flux events only occurred following the establishment of the negative shear and
 449 during the recovery of the mid-latitude USLM jet (associated with the “split” in the zonal
 450 mean).

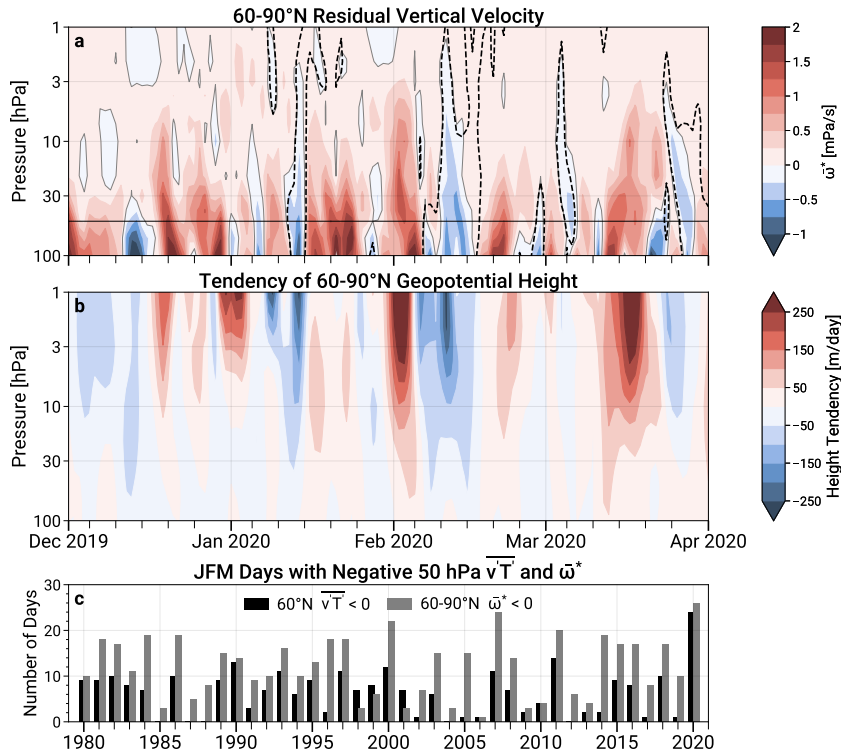


Figure 9. 60 – 90°N polar cap averaged residual vertical (pressure) velocity (a), the tendency of 60 – 90°N average geopotential heights (b), and the number of days with negative heat fluxes and a reversed residual circulation (c). The dashed contours in panel a show when the meridional eddy heat flux at 60°N was negative. Only pressure levels between 100 and 1 hPa are plotted in panels a and b. The black horizontal line in panel a corresponds to the 50 hPa level for which statistics are shown in panel c. Note that positive/negative pressure velocities indicate downward/upward motion, respectively.

451 The reflective zonal wind configuration and subsequent negative heat flux events
 452 aided in dynamically cooling and strengthening the polar vortex during the 2020 sea-

453 son. Figure 9 shows the 60-90°N average residual vertical pressure velocity ($\bar{\omega}^*$) and time
 454 tendencies of polar cap geopotential heights. The periods with negative heat fluxes at
 455 60°N are highlighted in Figure 9a by dashed contours. These events clearly correspond
 456 to reversals in the residual velocity that span almost the full polar stratospheric column.
 457 These events also coincide with negative 60-90°N polar cap height tendencies (Fig 9b).
 458 These polar cap height tendencies closely relate to changes in the thickness of the strato-
 459 spheric column, and the stratospheric NAM (which we have previously defined using 65-
 460 90°N polar cap heights), and thus the negative tendencies generally indicate the vortex
 461 cooled and strengthened during these events, consistent with prior studies (Shaw & Perl-
 462 witz, 2013, 2014; Dunn-Sigouin & Shaw, 2015). We further find that the 2020 JFM sea-
 463 son featured the largest number of days at 50 hPa with negative heat fluxes at 60°N and
 464 with a reversed polar cap residual vertical velocity in the MERRA-2 record (Fig 9c). Other
 465 years with large numbers of days with negative heat fluxes include 1989/1990, 1999/2000,
 466 and 2010/2011, which are all years that featured strong seasonal-mean polar vortices (see,
 467 e.g., Figure 3). However, 2019/2020 stands out even among these, having roughly dou-
 468 ble their number of days with negative heat fluxes. We also note that generally the win-
 469 ters having 10+ days with negative heat fluxes also featured one or more months with
 470 a split jet configuration in the zonal mean winds (not shown), similar to 2019/2020.

471 3.4 Polar Processing and Ozone Loss

472 The extremes in two-way wave coupling contributed to developing and maintain-
 473 ing a record strong polar vortex, which contributed to record ozone loss. Here we will
 474 show how characteristics of the polar vortex and conditions within it were conducive for
 475 the chemical destruction of ozone. We examine diagnostics of polar processing, and com-
 476 pare with other years with strong and cold polar vortices and/or large ozone loss, includ-
 477 ing 1996/1997 (Coy et al., 1997; Manney et al., 1997; Newman et al., 1997), 2010/2011
 478 (Manney et al., 2011), and 2015/2016 (Manney & Lawrence, 2016; Matthias et al., 2016).
 479 While the 2015/2016 winter did not culminate in a significant early-spring stratospheric
 480 ozone deficit, it did feature a very strong and unusually cold polar vortex that was cut
 481 short because of an early final warming. In this way, 2015/2016 serves as a foil to the
 482 other cases as an example of extreme polar processing conditions that did not lead to
 483 an extreme in stratospheric ozone.

484 The 2019/2020 polar vortex was exceptionally strong and long lived in the lower
 485 stratosphere, providing a robust containment vessel for chemical processing to occur in
 486 early spring as sunlight returned. Figure 10 shows time series of vortex area and max-
 487 imum potential vorticity (PV) gradients on the 490 K isentropic surface (around 50 - 60
 488 hPa). While the 2019/2020 vortex at 490 K was larger than normal in November, it was
 489 only about average size from December through January. However, the vortex remained
 490 at a roughly constant size between 20-25 million km² until the beginning of April, at which
 491 point its size was among the largest on record. In the lower stratosphere, strong PV gra-
 492 dients are known to inhibit mixing into and out of the vortex, and thus the magnitude
 493 of PV gradients describes how well the vortex edge acts as a barrier to transport (e.g.,
 494 Hoskins et al., 1985; Jukes & McIntyre, 1987; Scott et al., 2004). Here we show PV gra-
 495 dients as a function of equivalent latitude, which describe how closely contours of PV
 496 are spaced in an equivalent area coordinate system (see, e.g., Butchart & Remsberg, 1986).
 497 The daily maximum PV gradients (which generally occur at the polar vortex edge) over
 498 the 2019/2020 season started out near normal but became anomalously strong begin-
 499 ning in January before reaching all-time record highs in February through April (Fig 10c).
 500 The size of the lower stratospheric vortex during 2019/2020 remained above 10 million
 501 km² longer than any other previous year (Fig 10b), even 1996/1997, which had the largest
 502 vortex region from late March through the beginning of May. Similarly, the extended
 503 November-April 2020 mean maximum PV gradients were the largest in the MERRA-
 504 2 record (Fig 10d).

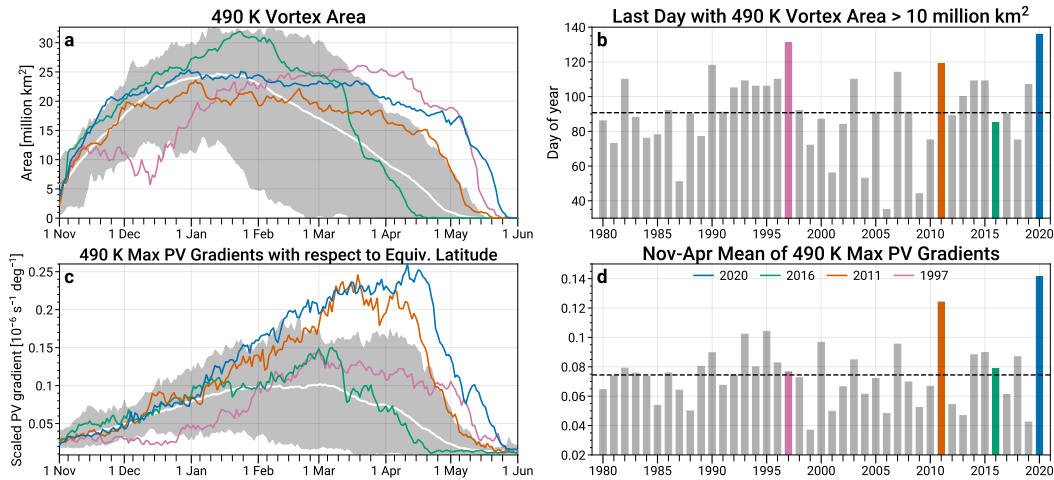


Figure 10. The left column shows daily time series of 490 K vortex area (a), and maximum PV gradients with respect to equivalent latitude (c). The right column shows derived statistics including the last day with 490 K vortex area above 10 million km² (b), and the November-March mean of the maximum PV gradients (d). The 2019/2020 season is highlighted in blue, with other relevant winters shown in green (2015/2016), orange (2010/2011) and pink (1996/1997). The grey envelopes and white lines in panels a and c represent (respectively) the climatological ranges and means after excluding the four highlighted years. The dashed horizontal lines in panels b and d represent the climatological average across the available years.

505 The 2019/2020 polar vortex was also the coldest in the MERRA-2 record for the
 506 formation of PSCs. In Figure 11, daily minimum temperatures at 50 hPa (Figures 11a)
 507 reached some all-time record lows in late November and early December, and temper-
 508 atures remained lower than the formation threshold for nitric acid trihydrate (NAT) PSCs
 509 until approximately March 25th. While this was not the latest date on record, 2019/2020
 510 still had the largest total number of days with temperatures below T_{NAT} (Fig 11b) be-
 511 cause of the early onset of the cold period. The vortex volume fraction of lower strato-
 512 spheric air with temperatures below T_{NAT} (V_{NAT}/V_{vort}) paints a consistent picture (Fig 11c);
 513 the 2019/2020 season attained all-time record maxima during some periods in mid-November
 514 and early December. Thereafter, the pool of cold air within the vortex remained rela-
 515 tively stable between fractions of 0.4 - 0.5 until early March (except for a brief dip in early
 516 February). Figure 11d suggests that roughly a third of the vortex volume in the lower
 517 stratosphere contained temperatures conducive to the formation of PSCs in the seasonal
 518 mean, the largest in any year in the MERRA-2 record.

519 Based on the results shown here, the 2019/2020 season had the greatest ozone loss
 520 *potential* ever observed. The polar processing conditions over the 2019/2020 season most
 521 closely resembled that seen during 2010/2011, which also had a relatively constant-sized
 522 vortex until late in the season, anomalously large PV gradients, and an extensive period
 523 of low temperatures. The 2015/2016 season also had an early onset of low temperatures
 524 and still holds some records for cold, but the vortex weakened much earlier in a dynamic
 525 final warming. The 1996/1997 season was effectively delayed by a month because an early
 526 winter warming kept the vortex small, weak, and warm, meaning less time was available
 527 for polar processing to occur.

528 Column ozone amounts in late winter and early spring suggest that exceptional chem-
 529 ical ozone loss did occur: Figure 12 shows the February-April (FMA) 2020 mean column

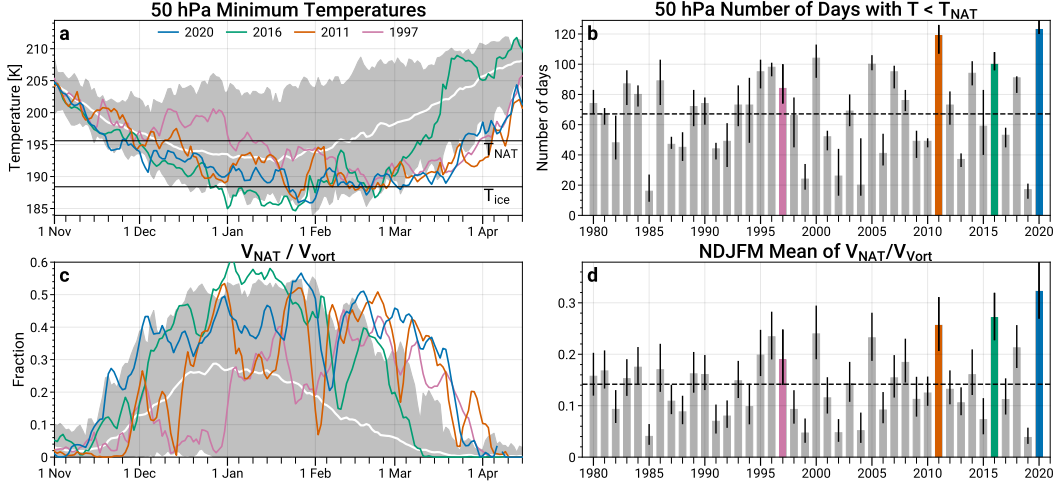


Figure 11. As in Figure 10, but the left column shows daily time series of 50 hPa minimum temperatures poleward of $40^\circ N$ (a), and the volume of air in the lower stratosphere with temperatures below the nitric acid trihydrate (NAT) polar stratospheric cloud (PSC) threshold (T_{NAT}) normalized by the vortex volume (V_{NAT}/V_{vort} ; c). The right column shows yearly integrated statistics, including the total number of days with temperatures below T_{NAT} at 50 hPa, and the November-March mean V_{NAT}/V_{vort} (d). Panel a has labeled horizontal black lines that represent the approximate formation thresholds for NAT and ice PSCs. The whiskers in panels b and d represent the ranges from accounting for ± 1 K uncertainties in the specific T_{NAT} threshold.

530 ozone anomalies alongside yearly time series of the FMA average of polar cap ($63 - 90^\circ N$)
 531 column ozone back to 1979 (the period over which regular total column ozone measure-
 532 ments were made by satellite instruments). Figure 12a shows that column ozone was anoma-
 533 lously low by more than 100 Dobson units (DU) over the pole for these three months.
 534 This ozone deficit is further reflected by the polar cap average time series shown in Fig-
 535 ure 12b, which shows that the 2020 FMA mean was the lowest on record since 1979, with
 536 a seasonal average less than 340 DU. The interpretation of low total column ozone amounts
 537 as they relate to chemical ozone depletion requires great caution, as dynamical influences
 538 related to tropospheric weather systems, lower stratospheric cold pools, and the loca-
 539 tion of the tropopause can cumulatively help to induce low column ozone amounts on
 540 daily to seasonal timescales (e.g., see discussions in Petzoldt, 1999; Manney et al., 2011).
 541 Reduced wave driving of the polar vortex and/or more frequent downward wave coupling
 542 events additionally lead to a weakened residual circulation that reduces the vertical re-
 543 supply of ozone, which can project onto anomalously low total column ozone amounts
 544 (Tegtmeier et al., 2008; Shaw & Perlwitz, 2014; Lubis et al., 2017). However, the com-
 545 bination of the persistent polar processing conditions conducive for chemical loss, and
 546 the persistently low column ozone values point to chemical depletion in 2019/2020 be-
 547 ing a large factor. Further, Manney et al. (2020) show evidence of chemical loss in vertically-
 548 resolved ozone profiles matching or exceeding that in 2011.

549 4 Discussion

550 We have provided a description of the unusual 2019/2020 polar vortex, and how
 551 it related to the observed climate extremes in the Arctic Oscillation and stratospheric
 552 ozone. Our results particularly highlight the important confluence of tropospheric and
 553 stratospheric conditions that overall made the exceptional polar vortex, AO, and ozone

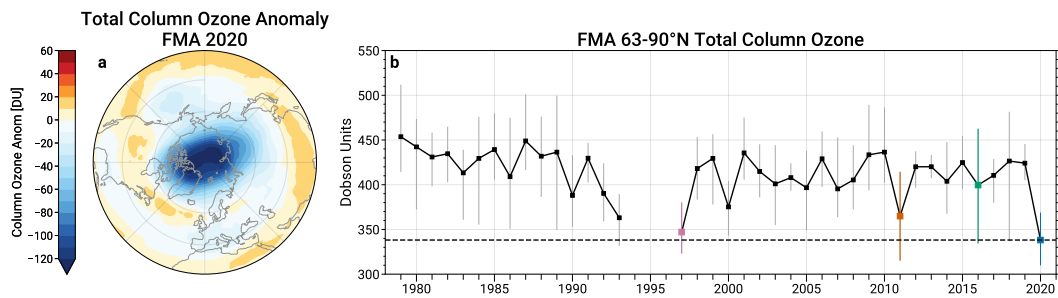


Figure 12. Map of Northern Hemisphere total column ozone anomalies averaged over February-April (FMA) 2020 (a) and yearly time series of the FMA mean 63-90°N polar cap ozone. The whiskers in panel b represent the range of the polar cap ozone values during the respective FMA seasons; the black dashed horizontal line is plotted at the mean value for 2020. The winters of 2019/2020, 2015/2016, 2010/2011, 1996/1997 are highlighted in the same colors as in Figures 10 and 11. The missing data between 1994-1996 is during a period without satellite column ozone observations.

554 depletion events possible. Together these events represent impacts of the most extreme
 555 and coherently coupled strong vortex event on the spectrum of observed Northern Hemi-
 556 sphere winters. There are a handful of previous winter seasons such as 1996/1997, 1999/2000,
 557 and 2010/2011 that were similar in nature to 2019/2020 in that they particularly involved
 558 anomalously strong, cold, and long-lived polar vortices (Figures 3 and 11), a large num-
 559 ber of negative heat flux days (Figure 9), and polar processing conditions more conducive
 560 for chemical ozone loss (Figure 12). However, these winters generally lacked the coher-
 561 ent coupling with the tropospheric circulation (Figures 3 and 5). In contrast, winters such
 562 as 1989/90 and 1992/1993 featured strong polar vortices, large numbers of negative heat
 563 flux days, and persistently positive tropospheric AO events, but lacked the unusually and
 564 persistently cold polar processing conditions necessary for exceptional chemical ozone
 565 loss (Figure 11). The fact that all these factors and events coincided in the same season
 566 of 2019/2020 makes it truly extraordinary.

567 Our paper provided a general overview of the extremes that occurred during the 2019/2020
 568 winter and how they developed. Further studies are necessary to fill in the details of mech-
 569 anisms, observations, predictability, and of the full range and magnitude of impacts. Be-
 570 low we pose some research questions motivated by the present work:

571 1. *What were the drivers (if any) of the strong vortex and/or AO events over internal*
 572 *variability?*

573 Interannual variability of the Arctic polar vortex is influenced by a variety of back-
 574 ground climate forcings and boundary conditions that act on sub-seasonal to seasonal
 575 timescales. These “drivers” impact the generation of waves in the troposphere, or influ-
 576 ence how they propagate through the atmosphere. Detailed modeling and attribution
 577 studies will be necessary to determine whether such processes played a role in the de-
 578 velopment of the strong polar vortex and/or the AO event over simple internal variabil-
 579 ity.

580 For example, sea surface temperatures (SSTs) in various regions have been linked
 581 to seasonal variability in the Arctic polar vortex. Some studies tied the previous strong
 582 and cold springtime polar vortices of 1997 and 2011 to positive SST anomalies in the north
 583 central Pacific (Hurwitz et al., 2011, 2012); more generally, SSTs in this region have been
 584 shown to modulate tropospheric planetary wave activity and the strength of the vortex

(e.g., Hu et al., 2018; Xie et al., 2020). Positive SST anomalies in the Indian Ocean have also been shown to encourage a strengthened Arctic polar vortex and positive NAM in the troposphere (Hoerling & Kumar, 2002; Hoerling et al., 2004; Li et al., 2010; Fletcher & Kushner, 2011), particularly in isolation from impacts by the El Niño-Southern Oscillation (ENSO) (Fletcher & Cassou, 2015). It is worth noting that the boreal autumn of 2019 featured a record strong Indian Ocean dipole (IOD) event (see, e.g., Johnson, 2020) and warm north Pacific SSTs from a marine heatwave (see, e.g., L’Heureux, 2019), amidst largely neutral ENSO conditions. A recent study by Hardiman et al. (2020) attributes predictability of the North Atlantic Oscillation (NAO) during winter 2019/2020 to this unusual IOD event, and particularly highlights the role of a stratospheric pathway related to a strengthened polar vortex. Other background forcings and boundary conditions that have been shown to impact the polar vortex include the tropical tropospheric Madden-Julian oscillation (e.g., Garfinkel, Feldstein, et al., 2012; Garfinkel et al., 2014; Liu et al., 2014; R. W. Lee et al., 2019), and the tropical stratospheric quasi-biennial oscillation (QBO; e.g., Baldwin et al., 2001; Garfinkel, Shaw, et al., 2012; White et al., 2016; Lubis et al., 2016; Lu et al., 2020). The QBO during the 2019/2020 winter was in the midst of a “disruption”, the second on record (Anstey et al., 2020), and it is presently unknown how such a disruption may have impacted the Arctic polar vortex during the season.

2. *How well were the strong polar vortex and AO events predicted by sub-seasonal to seasonal forecast models, and did the stratosphere contribute to tropospheric forecast skill?*

It is possible that some fraction of skill in sub-seasonal to seasonal (S2S) forecasts during the 2019/2020 winter and spring could be related to skill in predicting the strong polar vortex event, or being initialized with it. Studies have consistently shown a relationship between wintertime polar stratospheric initial conditions and improved S2S forecast skill (e.g., Sigmond et al., 2013; Tripathi, Baldwin, et al., 2015; Tripathi, Charlton-Perez, et al., 2015; Scaife et al., 2016; Nie et al., 2019). Recent work suggests there is also a relationship between model skill in predicting the stratosphere and skill for the troposphere (e.g., Domeisen et al., 2020a, 2020b). As mentioned above, a recent study by Hardiman et al. (2020) finds that the IOD conditions in late autumn/early winter influenced the strength of the polar vortex, which then impacted the NAO. Another recent study submitted for this special issue by S. H. Lee et al. (2020) found that ensemble members in a multi-model composite of seasonal forecasts that better predicted the strength of the 2019/2020 polar vortex also better predicted the anomalous tropospheric state.

A more complete accounting of the impacts related to stratosphere-troposphere coupling is also warranted: the reflective state of the stratosphere and multiple downward wave coupling events may have had a direct influence on tropospheric weather and circulation during the 2019/2020 winter and early spring. Downward wave reflection events have themselves been shown to help initiate positive phases of the North Atlantic Oscillation (Shaw & Perlwitz, 2013; Dunn-Sigouin & Shaw, 2015), and to occasionally directly induce weather events such as North Pacific blocking and cold spells in North America and Eurasia (Kodera et al., 2008; Kodera & Mukougawa, 2017; Matthias & Kretschmer, 2020).

3. *What were the relative roles of dynamical transport versus chemical loss processes in determining the low early spring column ozone?*

The anomalous polar cap ozone during the late winter and early spring of 2020 was clearly record breaking. The low ozone is generally consistent with the persistently strong polar vortex, which would have led to depressed ozone amounts due to a weakened residual circulation, and enhanced chemical loss due to the persistently cold polar vortex (Tegtmeier et al., 2008; Shaw & Perlwitz, 2014; Lubis et al., 2017). In 2010/2011 (the winter previously having the most extreme ozone loss) the individual contributions from transport

637 and chemical loss were both found to be record breaking based on a mixture of obser-
638 vations and models (e.g., Balis et al., 2011; Manney et al., 2011; Sinnhuber et al., 2011;
639 Adams et al., 2012; Strahan et al., 2013; Griffin et al., 2019). It will similarly be nec-
640 essary for studies to utilize a variety of observations and models to determine the rel-
641 ative roles of dynamical versus chemical impacts on low column ozone in spring 2020,
642 in addition to providing quantitative vertically-resolved chemical loss estimates. For ex-
643 ample, Manney et al. (2020, published in this special collection) use observations of rel-
644 evant chemical species from the Aura Microwave Limb Sounder to illustrate the chem-
645 ical and transport processes leading to exceptional chemical ozone loss and record low
646 ozone by spring 2020. Other studies presently submitted for this special collection and
647 elsewhere further explore the detailed evolution of ozone during the season using a vari-
648 ety of measurements and models (Dameris et al., 2020; Grooß & Müller, 2020; Inness
649 et al., 2020; Wohltmann et al., 2020), and more are in preparation.

650 *4. Were there downstream impacts related to the strong vortex, ozone deficit, and per-*
651 *sistent positive tropospheric AO events?*

652 The strong polar vortex, low ozone, and positive AO events that occurred in the
653 late winter/early spring of 2020 were each record breaking on seasonal timescales, and
654 as a result, there is a possibility they had farther-reaching consequences. For example,
655 it is possible that the depleted ozone into spring 2020 may have helped to maintain the
656 positive AO through April. One modeling study has shown that negative Arctic ozone
657 anomalies can cause a feedback on the strength of the vortex that increases the prob-
658 ability of a positive tropospheric AO (Karpechko et al., 2014), in a similar manner to
659 the observed tropospheric impacts of the Antarctic ozone hole (Thompson & Solomon,
660 2002; Shindell & Schmidt, 2004; Thompson et al., 2011). This kind of relationship be-
661 tween stratospheric ozone and the tropospheric circulation underpins why recent stud-
662 ies have suggested that springtime Arctic stratospheric ozone anomalies are linked with
663 surface temperatures and precipitation in specific regions for weeks to months ahead (e.g.,
664 Calvo et al., 2015; Ivy et al., 2017; Xie et al., 2018; Stone et al., 2019; Wang et al., 2020).

665 Additional climatologically relevant impacts are also possible: One recent study
666 illustrated that springtime stratospheric ozone intrusions are strongly impacted by the
667 abundance of ozone in the lowermost stratosphere in early spring (Albers et al., 2018),
668 meaning there could be a signature of the 2020 low ozone event in subsequent ozone in-
669 trusions of spring 2020. Another recent study has shown a relationship between a pos-
670 itive AO in the winter and early spring and increased fire activity and burn area in south-
671 eastern Siberia, a region where carbon release by fires can accelerate Arctic warming (Kim
672 et al., 2020). Yet another recent study has found a link between the timing of the spring-
673 time Arctic polar vortex breakdown and the distribution of sea ice thickness anomalies
674 all the way until the following autumn (Kelleher et al., 2020). Further study will be re-
675 quired to determine whether responses consistent with the above mentioned relationships,
676 or other events, arise due to influences from the exceptional 2019/2020 winter and spring.

677 These and other questions will be the focus of further work; we expect that many
678 will be addressed in the Journal of Geophysical Research/Geophysical Research Letters
679 Special Collection on the exceptional 2019/2020 Arctic polar vortex in which this arti-
680 cle appears.

681 **5 Conclusions**

682 The 2019/2020 NH stratospheric polar vortex was remarkably strong. The west-
683 erly stratospheric circulation represented by the polar vortex was the strongest on record
684 for December-March winter seasons back to 1979/1980; if considering earlier years back
685 to 1958/1959 for which data are more uncertain, 2019/2020 ranks among the top three,
686 although it depends on the specific level under consideration (e.g., 2019/2020 remains

687 the strongest at 100 hPa). The robust polar vortex appears to have developed due to
 688 a combination of weak tropospheric wave driving and a series of downward wave cou-
 689 pling events that occurred following the development of a reflective configuration of the
 690 polar vortex. Numerous aspects of the 2019/2020 winter and early spring were record
 691 breaking, and involved extremes in two-way troposphere-stratosphere coupling.

692 The positive AO and positive stratospheric NAM developed as a coherent event
 693 spanning the troposphere and stratosphere. As a result, the direction of causality be-
 694 tween the strongly positive NAM in the stratosphere and strongly positive AO in the
 695 troposphere is somewhat unclear. However, the persistence of the exceptionally strong
 696 vortex throughout the stratosphere suggests a stratospheric influence on the AO is more
 697 likely. Furthermore, downward wave coupling events are known to initiate tropospheric
 698 circulation anomalies consistent with a positive AO (Shaw & Perlwitz, 2013; Dunn-Sigouin
 699 & Shaw, 2015), meaning that the stratospheric wave reflection events that occurred dur-
 700 ing the 2019/2020 winter likely helped to maintain the positive AO. The January-March
 701 2020 mean AO was the largest on record and persistently positive. Large fractions of the
 702 observed surface temperature and precipitation anomalies in JFM were consistent with
 703 this large amplitude AO event, including a large portion of the record warmth that oc-
 704 curred over Eurasia.

705 The strong and long-lived polar vortex also provided ideal conditions for chemi-
 706 cal ozone destruction to take place. In the lower stratosphere, the polar vortex was a ro-
 707 bust transport barrier and very long lived, which isolated Arctic air during the key tran-
 708 sition period out of polar night. Furthermore, temperatures low enough to form polar
 709 stratospheric clouds within the vortex developed early in the season, and on average en-
 710 closed about a third of the vortex volume. In total, the number of days with such low
 711 temperatures exceeded 4 months. These conditions are unprecedented back to 1979/1980,
 712 making 2019/2020 the season with the greatest ozone loss potential on record. Polar cap
 713 column ozone amounts subsequently reached low levels never before observed in the Arc-
 714 tic at this time of year.

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 726 data and statistics from OMPS and other instruments are compiled and made available
 727 via NASA's OzoneWatch resource at <https://ozonewatch.gsfc.nasa.gov/data/>.
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