

Characteristics, Evolution, and Formation of Cold Air Outbreaks in the Great Plains of the United States

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ABSTRACT: Wintertime cold air outbreaks (CAOs) in the Great Plains of the United States have significant socioeconomic, environmental, and infrastructural impacts; the events of December 1983 and February 2021 are key examples of this. Previous studies have investigated CAOs in other parts of North America, particularly the eastern United States, but the development of CAOs in the Great Plains and their potential subseasonal-to-seasonal (S2S) predictability have yet to be assessed. This study first identifies 37 large-scale CAOs in the Great Plains between 1950 and 2021, before examining their characteristics, evolution, and driving mechanisms. These events occur under two dominant weather regimes at event onset: one set associated with anomalous ridging over Alaska and the other set associated with anomalous pan-Arctic ridging. Alaskan ridge CAOs evolve quickly (i.e., on synoptic time scales) and involve stratospheric wave reflection. Conversely, Arctic high CAOs are preceded by weak stratospheric polar vortex conditions several weeks prior to the event. Both categories of CAOs feature anomalous upward wave activity flux from Siberia, with downward wave activity flux over Canada seen only in the Alaskan ridge CAOs. The rapid development of the Alaskan ridge CAOs, also linked with a North Pacific wave train and anomalous wave activity flux from the central Pacific, suggests that these events could be forced by tropical modes of variability. These findings present evidence that different forcing mechanisms, with contrasting time scales, may produce distinct sources of predictability for these CAOs on the S2S time scale.

KEYWORDS: Extreme events; Stratosphere-troposphere coupling; Winter/cool season; Subseasonal variability

1. Introduction

Wintertime cold air outbreaks (CAOs) are extreme cold events of temporally and spatially large scales that affect large portions of the midlatitudes in the winter, through a displacement of air southward from a polar origin (Cellitti et al. 2006; Wheeler et al. 2011; Smith and Sheridan 2018a, 2020). CAOs cause noteworthy socioeconomic, environmental, and infrastructural impacts (Analitis et al. 2008; Smith and Sheridan 2019). In the United States, there were an average of 27 cold-related fatalities per year between 1991 and 2020 (NWS 2021b). Additionally, freeze events in the United States have cost \$3.5 billion per event between 1981 and 2020 (NCEI 2021).

The Great Plains of the United States—the region that is east of the Rocky Mountains but west of the Mississippi River (Rossum and Lavin 2000; Kloesel et al. 2018)—has endured some of the strongest cold waves in North America (Fig. 1, black box). The 2-week-long extreme CAO that occurred in the Great Plains during December 1983 resulted in Oklahoma City's December average temperature being 13.2°F colder than average, leading to the freezing of water pipes and the main water supply, with many people left without water for days (McManus 2009). More recently, an extreme CAO impacted the Great Plains in February 2021, with the southern

states being the hardest hit. All of Texas, Oklahoma, and Arkansas were under winter storm warnings (NWS 2021a), while frozen gas lines and surging heating demand led to widespread power outages in Texas, along with numerous cold-related fatalities and road incidents (BBC News 2021). Understanding the causes and evolution of these events will aid progress in improving their predictions on the subseasonal-to-seasonal (S2S) time scale, defined by a lead time of two weeks to two months (Vitart and Robertson 2018).

Various studies have investigated the properties and precursors of CAOs in the United States (e.g., Konrad and Colucci 1989; Cellitti et al. 2006; Xie et al. 2017; Smith and Sheridan 2018a,b), though with a strong focus on CAOs in the eastern half of the United States. Anomalous ridging over Alaska is one recurring atmospheric signature for CAOs across North America, but mainly for northern and eastern regions (Konrad and Colucci 1989; Cellitti et al. 2006; Xie et al. 2017; Lee et al. 2019). However, there are also links between cold anomalies in the U.S. Great Plains with high-latitude anomalous ridging across the Arctic (e.g., Stan and Straus 2007; Lee et al. 2019). Thus, the development of extreme cold events in the United States is aided by the formation of large-scale patterns that help to transport Arctic air masses southward (Grotjahn et al. 2015), in association with overlying negative 500-hPa geopotential height (GPH) anomalies and upstream blocking (Loikith and Broccoli 2012; Westby and Black 2015; Xie et al. 2017). Furthermore, Cellitti et al. (2006) found that a weak tropospheric polar vortex, with increased Arctic ridging, is favored prior to CAO onset. This finding links to the conclusions of Smith and Sheridan (2018b), who showed that the tropospheric polar vortex and the stratospheric polar vortex (SPV) are persistently weak

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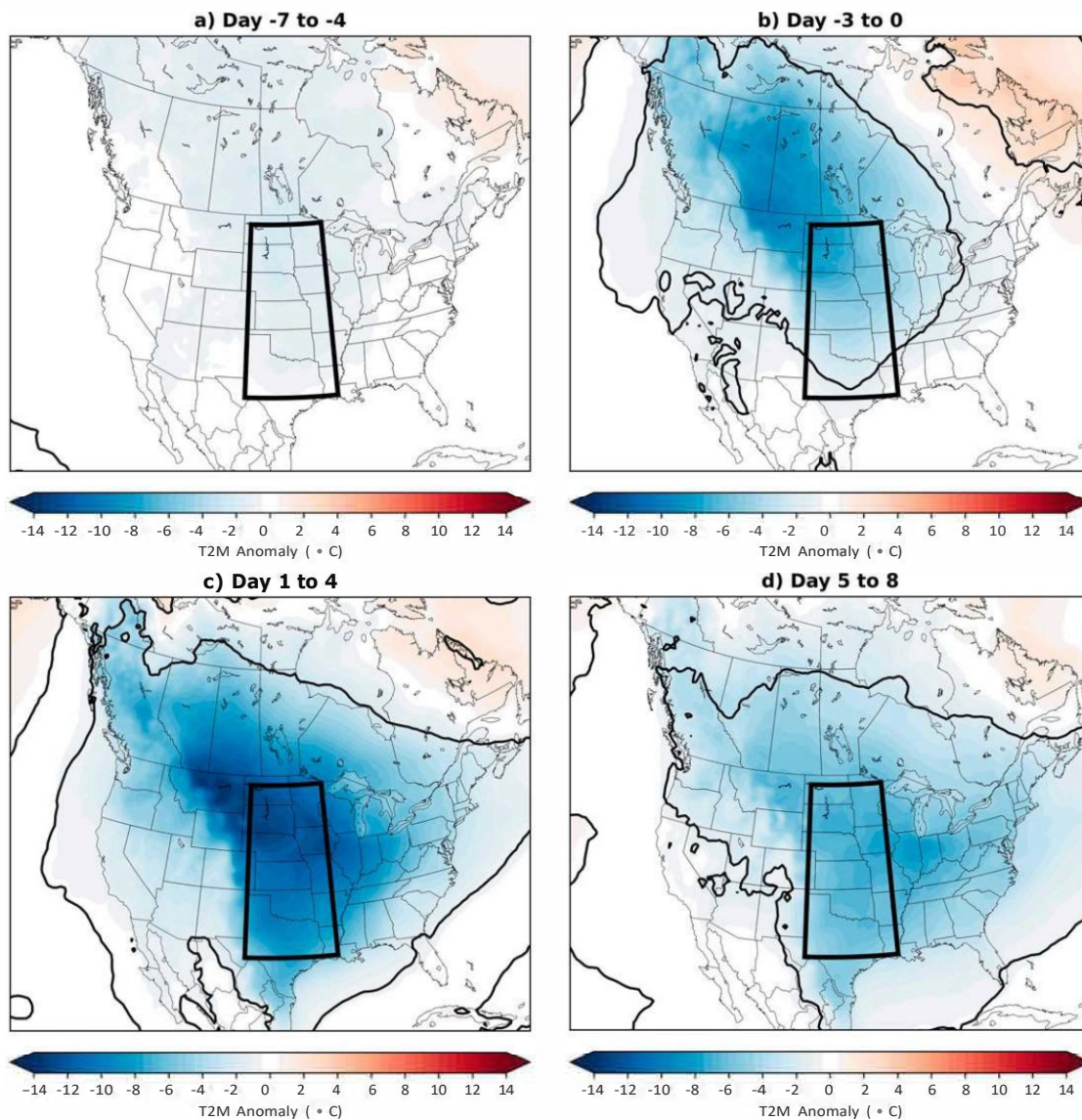


FIG. 1. Lagged composites of T2M anomalies ($^{\circ}\text{C}$; shaded) across North America for the 37 Great Plains CAOs, with respect to CAO onset (day 0) (a) from days -7 to -4, (b) from days -3 to -0, (c) from days +1 to +4, and (d) from days +5 to +8. Black contours represent significance at the 95th percentile level via a bootstrapping approach of 5000 iterations. The black box represents the region used to define the CAOs. Negative lags indicate days prior to CAO onset (day 0), and positive lags indicate days after CAO onset.

prior to eastern U.S. CAOs, leading to the negative phase of the northern annular mode (NAM; e.g., Thompson and Wallace 2000) and the North Atlantic Oscillation (NAO; e.g., Wallace and Gutzler 1981).

SPV variability can have significant impacts on surface weather (e.g., Baldwin and Dunkerton 2001; Limpasuvan et al. 2005; Domeisen et al. 2013; Afargan-Gerstman and Domeisen 2020), particularly following *major* sudden stratospheric warmings (SSWs). During these events, the dominant westerlies around the SPV reverse to easterlies in response to strong upward and poleward wave flux transporting westward angular momentum and anomalous heat, destroying the vortex (Baldwin and Dunkerton 2001; Nakagawa and Yamazaki 2006; Charlton and Polvani 2007). Weak SPV conditions then

enable downward propagation of easterlies, inducing the negative phase of the NAM into the troposphere (e.g., Thompson and Wallace 2000; Baldwin and Dunkerton 2001; Hitchcock and Simpson 2014; White et al. 2019) and an equatorward shift in the polar jet stream (Limpasuvan et al. 2004; Maycock et al. 2020). This downward coupling of the NAM in Eurasia and North America increases the probability of CAOs there (Kolstad et al. 2010), with North American CAOs favored under displaced SPV events (Mitchell et al. 2013).

However, the link between SPV variability and North American CAOs is less straightforward than the link between SPV variability and Eurasian CAOs. Alternatively, downward coupling via stratospheric wave reflection has been proposed to explain the stratospheric influence on CAO development

in North America. Rather than the planetary-scale wave absorption during weak SPV events, wave activity from the troposphere is reflected downward from the stratosphere, inducing tropospheric circulation changes (e.g., Perlwitz and Hamik 2003, 2004; Kodera et al 2008, 2013; Harnik 2009; Shaw et al 2010; Kretschmer et al 2018a; Matthias and Kretschmer 2020). Indeed, stratospheric wave reflection played a role in two North American cold spells during the winter of 2017/18 (Matthias and Kretschmer 2020). This mechanism requires two key conditions: 1) a high-latitude vertical reflecting surface in the upper stratosphere marked by negative wind shear and increased curvature of the zonal wind profile; and 2) a meridional waveguide in the middle to lower extratropical stratosphere, marked by a zonal wind maximum with increased curvature of the zonal wind in the meridional direction, i.e., with latitude (Perlwitz and Hamik 2003; Matthias and Kretschmer 2020).

Prompted by the Great Plains CAO in February 2021 and previous literature, this study investigates the general characteristics of CAOs in the Great Plains, what their dominant onset weather regimes are, the role that the stratosphere has in their development, and implications for S2S predictability. We hypothesize that Arctic ridging will be important for the development of Great Plains CAOs, along with Alaskan–Arctic ridging, as seen for other part of North America. Each of these high-latitude ridging areas likely exhibit unique connections to the stratosphere (Lee et al. 2019), in regard to Great Plains CAOs. Our results further emphasize the importance of stratosphere–troposphere dynamical coupling in winter time extreme weather and in S2S forecasts.

2. Data and methods

a. Data

This study uses daily averaged November–March (NDJFM) European Centre for Medium-Range Weather Forecasts (ECMWF) fifth reanalysis (ERAS) data (Hersbach et al. 2020), from 1 January 1950 to 31 March 2021 at a 0.5° horizontal resolution with 37 vertical levels from 1000 to 1 hPa. We use this dataset since it is the latest ECMWF reanalysis extending back to 1950. We analyze the following variables: 2-m temperature (TIM), GPH, zonal wind (u), meridional wind (v), and air temperature (T). From these, derived quantities are calculated. The first of these is the Plumb wave activity flux (WAF), a measure of quasi-stationary wave propagation as defined in Eq. (7.1) by Plumb (1985):

$$F_s = p \cos \phi \begin{pmatrix} v^2 - \frac{1}{2f \sin 2\phi} \frac{a(v'')^2}{a_z} \\ -u'v' + \frac{1}{2f \sin 2\phi} \frac{a(u'')^2}{a_z} \\ \frac{2f \sin \phi}{S} [v'T' - \frac{1}{2f \sin 2\phi} \frac{a(T'\phi')}{a_z}] \end{pmatrix}, \quad (1)$$

where p is pressure/1000 hPa, a is the radius of Earth, ϕ is geopotential, λ is longitude, ϕ is latitude, and S is the static stability, defined as

$$S = \frac{\partial \hat{T}}{\partial z} + \frac{\kappa \hat{T}}{H}, \quad (2)$$

where z is height, κ is the ratio of the gas constant for dry air to the specific heat at constant pressure, $H \approx 8300$ m is the scale height approximation for the atmosphere. Carats ($\hat{\cdot}$) denote the areal average north of 20°N, and primes ($'$) indicate departures from the zonal mean. Since planetary-scale waves are dominant in forcing stratospheric variability, the

WAF is filtered for wavenumbers 1–3 to capture their impact. Second, we calculate the 100-hPa meridional eddy heat flux ($v'T'$), used for capturing potential regions of enhanced vertical wave propagation within the stratosphere [note that the vertical component of F is proportional to $v'T'$ in (1)]. Finally, the quasigeostrophic streamfunction (ψ) is calculated by

$$\psi = \frac{Zg}{f}, \quad (3)$$

where Z is the GPH, $g = 9.81 \text{ m s}^{-2}$ is gravitational acceleration, $f = 2\Omega \sin \phi$ is the Coriolis parameter, with Ω being the angular velocity of Earth. Unless stated otherwise, all anomalies are calculated with respect to the daily 1981–2010 climatology, where we remove the seasonal cycle from each day.

b. Defining the cold air outbreaks

The Great Plains CAOs are defined by using the December, January, and February (DJF) linearly detrended and area-averaged TIM anomalies over the region of 92°–103°W and 3°N–48°N (i.e., the region defining the Great Plains). This region is outlined by the black box in Fig 1. CAOs meet both magnitude and temporal criteria: 1) magnitude: the magnitude of the daily TIM anomaly must be below the 10th percentile of the linearly detrended and area-averaged

2) DJF TIM anomalies.

Each CAO is separated from the end of a previous CAO by at least four days, as a measure of synoptic independence.

The start of a CAO (onset day) is the first day in which the above conditions are met. A relaxation criterion is also applied, in which a maximum of 1 day during a CAO is allowed to miss the magnitude threshold. This day cannot be the start or end day of a CAO, and the end of the CAO is the last day the magnitude threshold is met before two or more consecutive days miss the threshold. This relaxation captures events that may otherwise miss the criteria above, but are still likely to be impactful.

c. Composite analyses and weather regime identification

To diagnose the characteristics and driving mechanisms behind the CAOs, we produce a series of time-lagged com-

posite analyses using the variables mentioned earlier. While compositing is a powerful tool, a disadvantage is that averaging all the events masks event-to-event variability. Thus, to

complement the composites, we expand the k-means clustering method of Lee et al. (2019) for calculating weather regimes over North America (180° – 330° E and 20° – 80° N) back to 1950. This method is a form of unsupervised machine learning that groups patterns together that show similarities in a meteorological field. An empirical orthogonal function (EOF) analysis is performed on the linearly detrended NDJFM 500-hPa GPH anomalies, with respect to the 1979–2020 climatological mean to remain consistent with Lee et al. (2019). The data are weighted by the square root of the cosine of latitude prior to eigenanalysis. The 12 leading principal component (PC) time series are retained, accounting for approximately 80% of the explained variance. Five regimes are then defined, by using k-means clustering on the 12 PC time series via the Python package "sci-kit learn" (Pedregosa et al. 2011). We use five clusters, instead of four as in Lee et al. (2019), to account for any shifts on the decadal or longer time scale in the North Pacific over this longer period. Each day in the record is assigned a distinct regime based on the minimum Euclidean distance to the cluster centroid. The CAO events are then subsetted by the regime on day 0 (i.e., the first day of the CAO). Composites of relevant atmospheric variables are subsequently computed to assess the dynamics and evolution of the CAOs associated with different regimes. Statistical significance for all composite maps is computed by using bootstrap resampling of 5000 iterations, without replacement.

d. Examining stratospheric wave reflection

We also examine the role of wave propagation and stratospheric wave reflection on Great Plains CAOs. For wave reflection, we use the reflective index defined by Matthias and Kretschmer (2020). This reflective index (RI) describes the difference between area-averaged standardized anomalies of the 100-hPa ν' T' over Siberia and Canada (defined between latitudes 45° – 75° N and longitudes of 120° – 185° E and 225° – 250° E, respectively), as per Eq. (2) in Matthias and Kretschmer (2020):

$$RI = (\nu' T')_{Sib}^* - (\nu' T')_{Can}^* \quad (4)$$

where the asterisk (*) denotes standardized quantities and other variables are as defined earlier. A positive **RI** indicates more positive anomalous $\nu' T'$ in Siberia than in Canada which is more indicative of a wave reflection mechanism especially if there are positive anomalies in Siberia and negative anomalies in Canada (indicating anomalously downward propagating/possibly reflected waves).

To identify reflective surfaces in the stratosphere, the curvature (a) of the vertical and meridional profiles of the wnal wind are calculated via Eq. (1) in Matthias and Kretschmer (2020):

$$-\frac{\partial^2 u}{\partial x^2} \quad (5)$$

where x represents either height (z) or meridional distance (y). Increased curvature is defined by exceeding the climatological curvature.

3. Results

a. Characteristics of Great Plains CAOs

1) TEMPERATURE PATTERNS AND THE ATMOSPHERIC CIRCULATION

The definition for the Great Plains CAOs yielded 37 events between 1950 and 2021. Figure 1 shows the lag composites for the T2M anomalies across North America, where day 0 corresponds to the onset day of the CAOs. From days -7 to -4, there is a lack of significant ($p < 0.05$) TIM anomalies across the Great Plains (Fig. 1a). From days -3 to 0 (Fig. 1b), significant negative TIM anomalies appear across the Great Plains, with the coldest anomalies across the northern portion of the domain. The extent and magnitude of cold anomalies significantly increase across the region of study after day 0 (Fig. 1c), with widespread negative TIM anomalies ranging from -6° C in Texas to around -13° C in the northern Great Plains. Even colder TIM anomalies exist to the northwest of the Great Plains, indicating the dominant northwest to southeast propagation. Between days +5 and +8, the magnitude of TIM anomalies decreases and becomes more uniform in the Great Plains, ranging from approximately -4° to -7° C (Fig. 1d). The presence of these significant, but weaker, TIM anomalies in Fig. 1d demonstrates the long duration of some of these events.

The TIM composites in Fig. 1 show negative anomalies building across the Great Plains from the northwest, which is likely associated with large-scale circulation patterns. Figures 2 and 3 show the 500- and 50-hPa lagged composites, respectively, of statistically significant linearly detrended GPH anomalies (shaded) and the full geopotential height field (contours) across the Northern Hemisphere for all CAO events. Days -11 through -8 show the development of a dipole of 500-hPa GPH anomalies in the North Atlantic (Fig. 2a), resembling the negative phase of the NAO, and significant, positive 500-hPa GPH anomalies develop over Alaska from days -7 to -4 (Fig. 2b). From days -3 to 0, a region of positive 500-hPa GPH anomalies exists across Greenland, but a stronger, anomalous ridge rapidly develops over Alaska (Fig. 2c), which is a well-known pattern favorable for U.S. CAOs (e.g., Cellitti et al. 2006; Xie et al. 2017; Lee et al. 2019). At the same time, significant anomalous troughing develops over North America, accompanying cold air advection across the Great Plains via a north to northwesterly flow (Fig. S1 in the online supplemental material). The anomalous ridging over Alaska persists through to days +5 to +8 (Figs. 2d,e), while the anomalous troughing over North America weakens and pushes eastward (Figs. 2e,f).

In the stratosphere, the 50-hPa **GPH** anomaly composites show a reasonably equivalent-barotropic pattern relative to the 500-hPa GPH anomalies, especially in the days leading to CAO onset (Figs. 3a-c). However, significant negative GPH anomalies over North America in the stratosphere persist longer than those in the troposphere. Positive 50-hPa GPH anomalies are also present over Eurasia, alongside significant anomalous troughing over North America, indicative of an elongated SPV over North America (Figs. 3d-f), whereby the

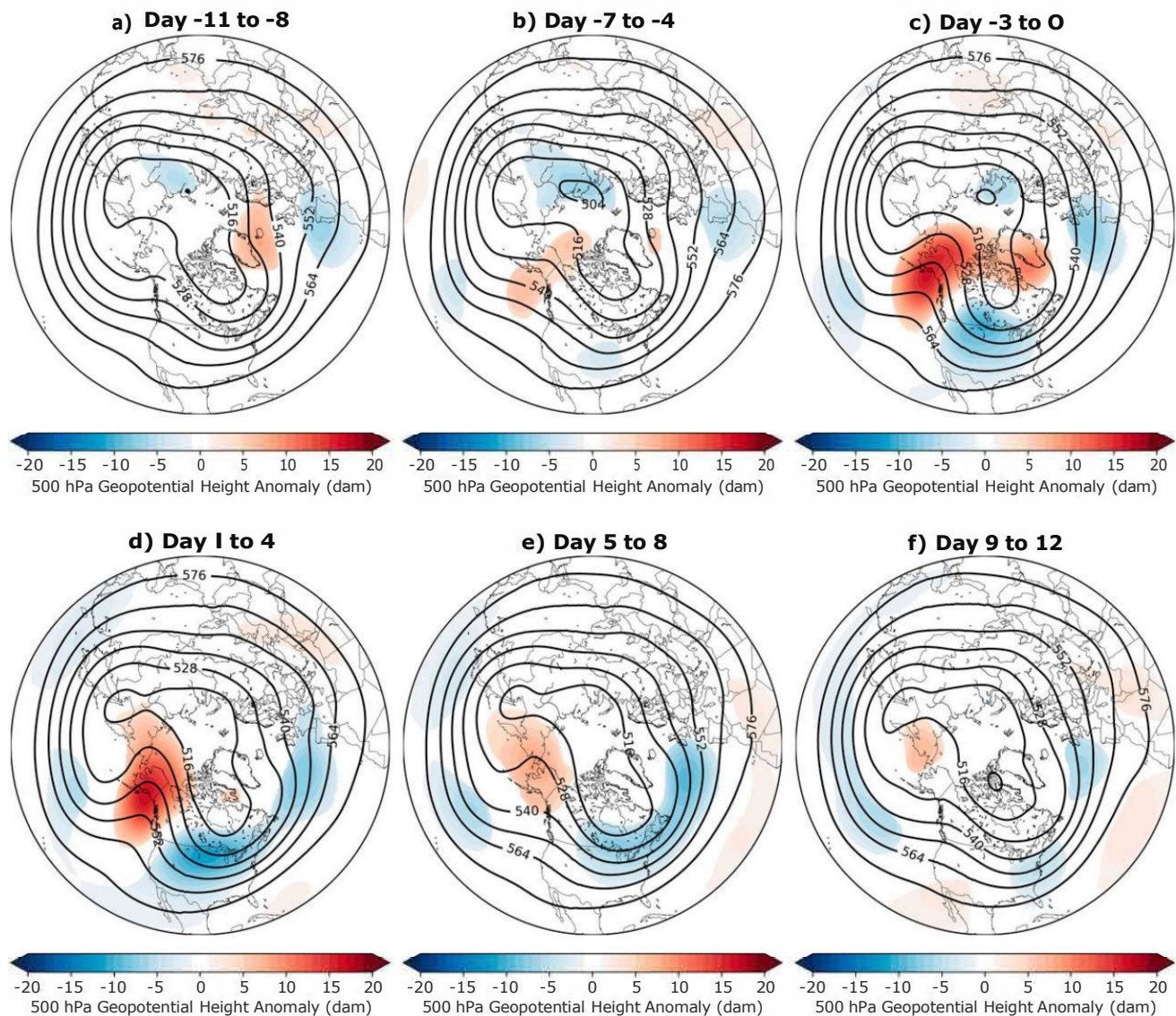


FIG. 2. Lagged composites of the full-field GPH (dam; black contours) and only the significant ($p < 0.05$) linearly detrended 500-hPa GPH anomalies (dam; shaded) across the Northern Hemisphere for the 37 Great Plains CAOs, with respect to CAO onset (day 0) at 4-day intervals from (a) days -11 to -8 through (f) days +9 to +12. Statistical significance is calculated via a bootstrapping approach of 5000 iterations. Negative lags indicate days prior to CAO onset day (day 0), and positive lags indicate days after CAO onset.

shape of the vortex is more elliptical than circular in nature (Cohen et al. 2021).

2) NORTH AMERICAN WINTER WEATHER REGIMES AND GREAT PLAINS CAOs

In the previous section, we found that there is anomalous tropospheric ridging in two key areas for Great Plains CAOs on and before CAO onset: Alaska and Greenland. Thus, we next investigate which CAOs are associated with either (or both) of these dominant patterns, to assess potential differences in dynamics and predictability. Our method of choice is to conduct regime analysis, like Lee et al. (2019) but for the extended period 1950–2021. The regime calculation generated five North American weather regimes: the Alaskan ridge (**AKR**), the

Arctic high (ArH), the Pacific trough (PT), the West Coast ridge (WCR), and the Arctic low (ArL). These regimes are shown in Figs. 4a–e, respectively. We find the same AKR, ArH, PT, and ArL regimes as those generated in Lee et al. (2019) but with a differing order of frequency, with the WCR as the additional regime. Our analysis has the following regimes from most to least frequent: the ArL (Fig. 4e, 24.0%) with anomalous North Pacific ridging and anomalous troughing in the Arctic regions, the WCR (Fig. 4d, 21.6%), which features a 500-hPa GPH anomaly wave train in the Pacific–North American (PNA) region, the PT (Fig. 4c, 20.1%), which shows an enhanced Aleutian low, the ArH (Fig. 4b, 17.3%) with anomalous ridging over Greenland and northern Canada, and the AKR (Fig. 4a, 17.0%) with anomalous ridging over Alaska.

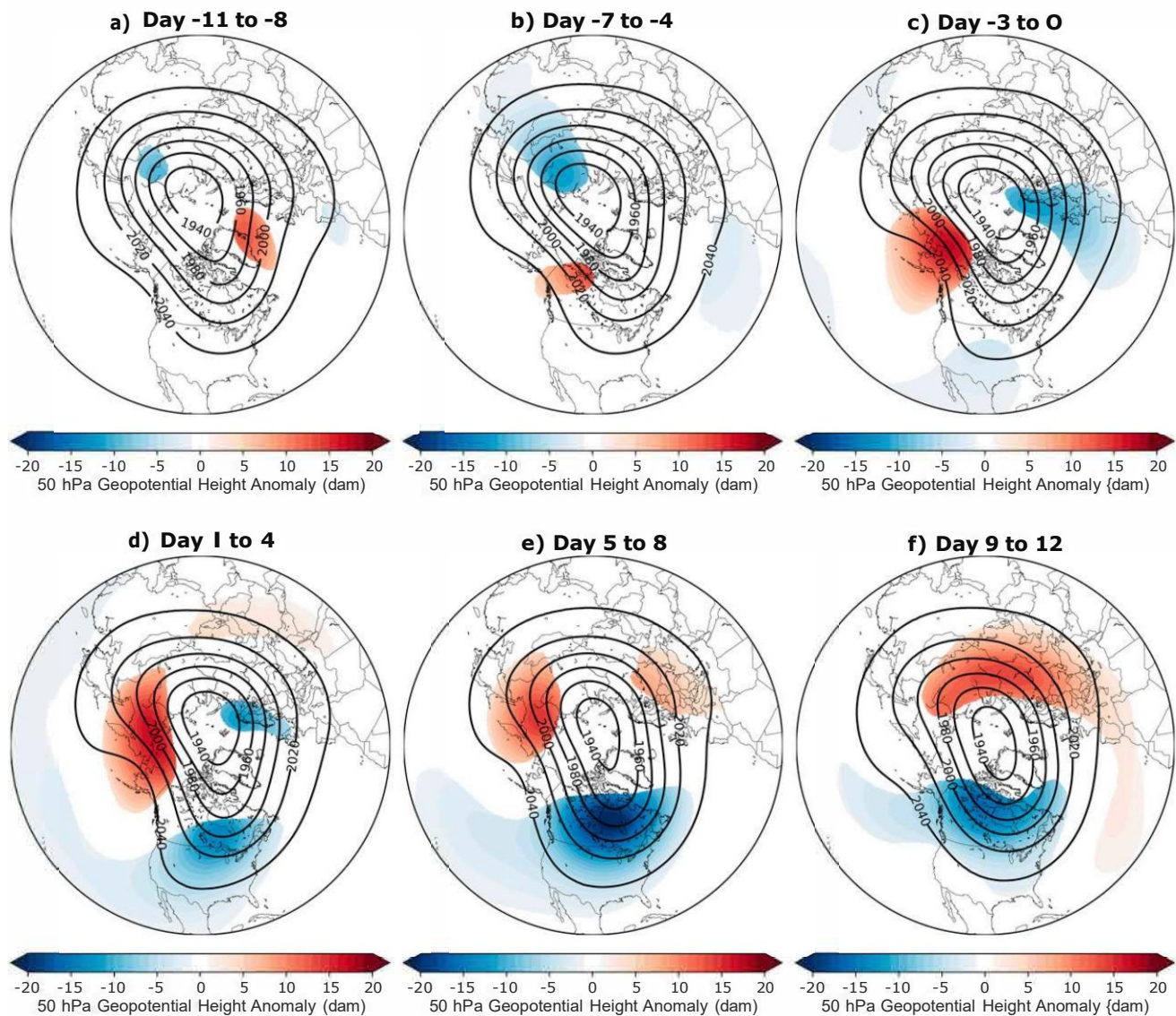


FIG. 3. As in Fig. 2, but for 50-hPa full-field GPH (dam, contours) and only significant ($p < 0.05$) linearly detrended GPH anomalies (dam; shading).

After assigning each day in NDJFM from 1 January 1950 to 31 March 2021 a weather regime classification, we next subset the 37 Great Plains CAOs based on the regime on the CAO onset day. Of the 37 Great Plains CAOs, 12 are AkR-CAOs, 18 are ArH-CAOs, and 7 are WCR-CAOs. Unsurprisingly, no CAOs are defined as PT- or ArL-CAOs, as these regimes are not conducive for cold air advection into the Great Plains (Figs. 4c,e). Figure 5 shows CAO onset day composites of 500-hPa linearly detrended GPH anomalies for the AkR- (Fig. 5a), ArH- (Fig. 5b), and WCR-CAOs (Fig. 5c). Qualitatively, the AkR- and ArH-CAO composites match to their characteristic patterns in Figs. 4a and 4b exceptionally well, but with larger anomalies, suggesting that these CAOs develop from amplified tropospheric patterns. The large sample size used to generate the characteristic regime composites could also result in comparatively smaller magnitude GPH anomalies

to that of the CAOs (Fig. 5). Quantitatively, these patterns also match well with their corresponding regime patterns across the North American domain; the AkR regime (Fig. 4a) correlates with the AkR-CAO composite (Fig. 5a) with a value of $r = 0.86$ ($p < 0.05$), and the ArH regime (Fig. 4b) correlates with the ArH-CAO composite (Fig. 5b) with a value of $r = 0.89$ ($p < 0.05$). The WCR-CAO composite pattern (Fig. 5c) correlates less strongly to its characteristic pattern (Fig. 4d); however, it is still statistically significant ($r = 0.70$; $p < 0.05$). This weaker pattern correlation arises because "WCR-CAOs" do not feature anomalous ridging across the entire western United States (Fig. 5c). Our regime-based composites (Figs. 5a,b) demonstrate that the composites for all CAOs (Fig. 2c) mask the relative importance of the AkR and ArH regimes which may involve very different characteristics and evolution, which we next explore.

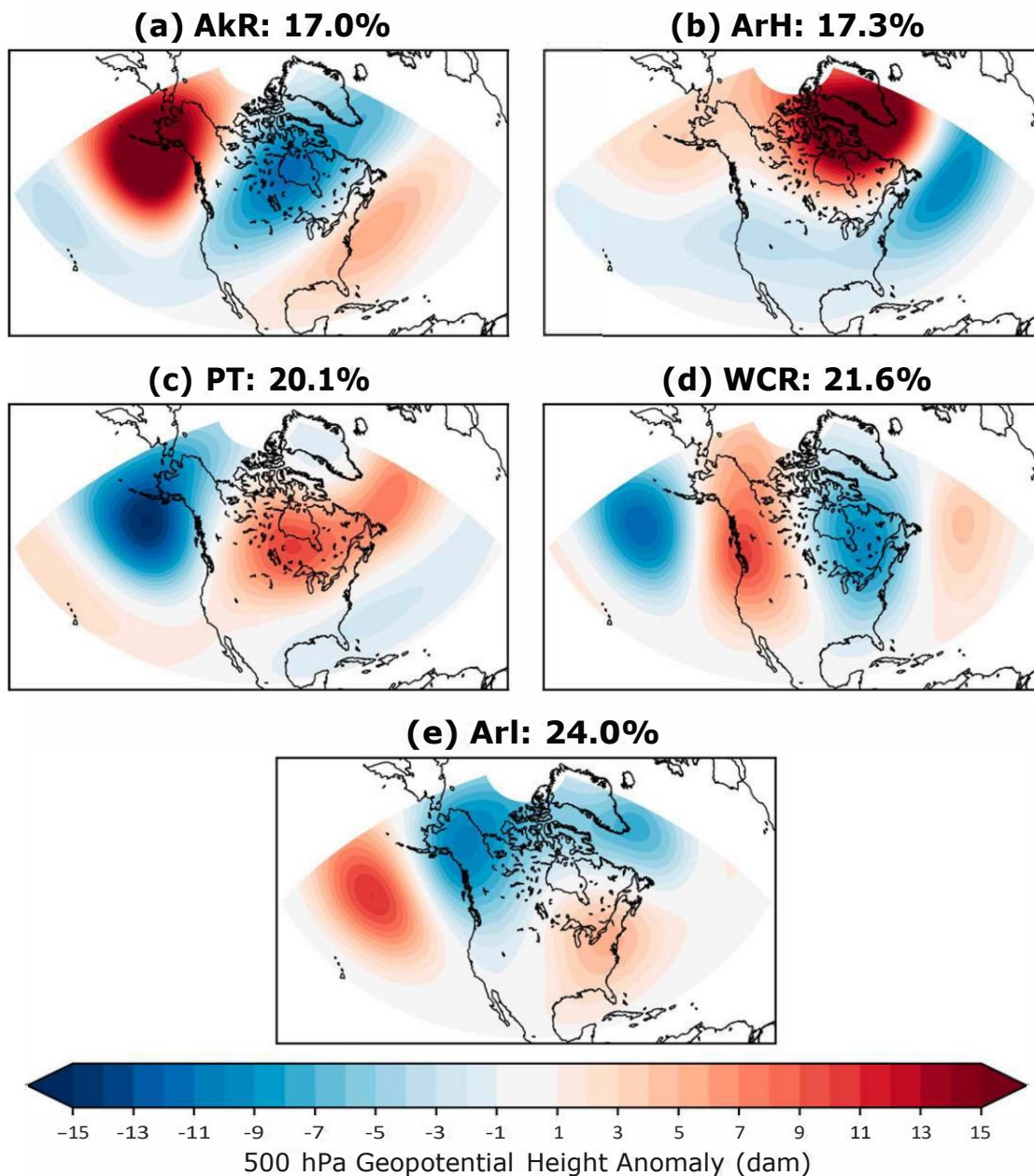


FIG. 4. Composite 500-hPa linearly detrended GPH anomalies (dam; shaded), for the five North American weather regimes for NDJFM between 1950 and 2021: (a) Alaskan ridge (AkR), (b) Arctic high (ArH), (c) Pacific trough (PT), (d) West Coast ridge (WCR), and (e) Arctic low (ArL). Percentages show the frequency of occurrence of a given regime throughout 1950–2021.

3) MAGNITUDE AND LONGEVITY OF THE COLD AIR OIjfbREAKS

To further analyze the characteristics of the three CAO types, Fig. 6 shows the magnitude of the CAO events, found by taking the mean domain-averaged TIM anomalies over the duration of the event (noted in bold in Fig. 6). The average duration of a CAO is 7.5 days, the average magnitude is -10.4°C , and a CAO occurs approximately once every two winter seasons. Exceptions are two CAOs in the winter

seasons of 1972/73, 1978/79, and 1983/1984, and four CAOs in 1977/1978. The intense and persistent AkR-CAO on 16 December 1983 stands out, with an average TIM anomaly of -13°C over 15 days. This event was an exceptional, and prolonged, CAO across much of North America (e.g., McManus 2009). Figure S2 illustrates the severity of the CAO events by using a "CAO severity index," i.e., the product of the magnitude and duration of each event. The three most severe CAOs based on this index are 1) 16 December 1983 (AkR, -195°C day), 2) 7 February 2021 (ArH, -177°C day), and 3) 21 January

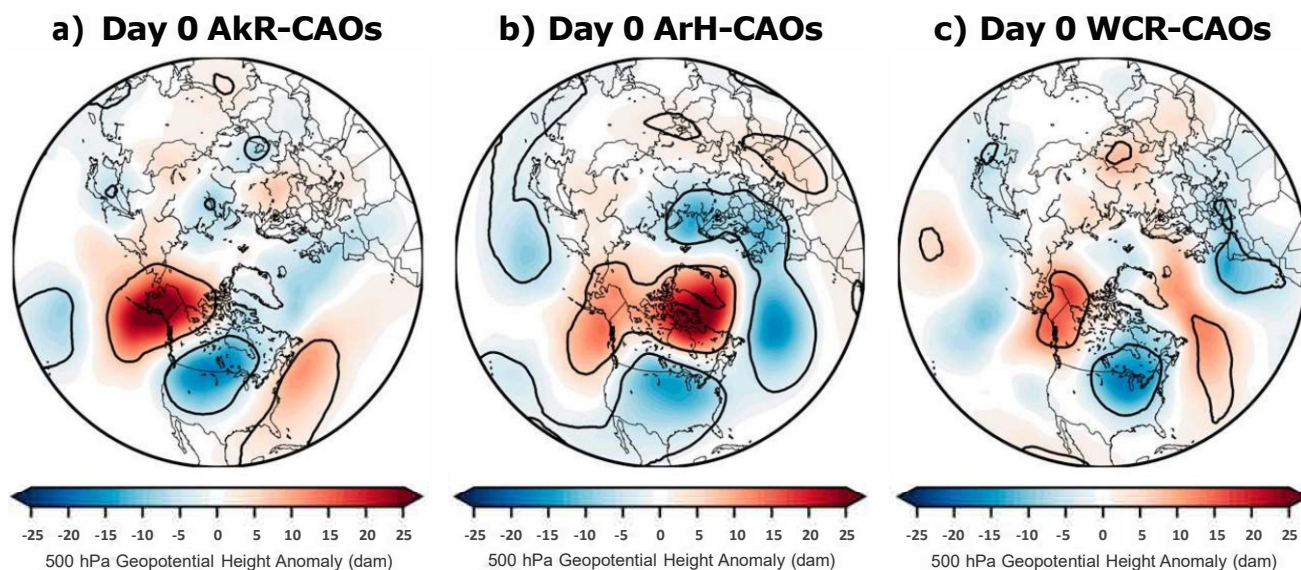


FIG. 5. Onset day composites of linearly detrended 500-hPa GPH anomalies (dam; shaded), with respect to the 1981–2010 climatology, are shown for (a) AkR-, (b) ArH-, and (c) WCR-CAOs. Black contours represent statistical significance ($p < 0.05$) via 5000-iteration bootstrapping without replacement.

1966 (ArH, -148°C day). Table 1 shows mean statistics for Great Plains CAOs by regime type. ArH-CAOs are slightly longer on average, the magnitude of the AkR-CAOs is slightly greater, and the WCR-CAOs are shorter and less intense than the ArH- and AkR-CAOs (Fig. 6 and Table 1), but these differences are not statistically significant. Given these three regime types have anomalous ridges in the PNA and North Atlantic sectors, their related CAO events may have different wave forcing, stratospheric influences, and developmental time scales.

b. Wave activity and the influence of the stratosphere on Great Plains CAOs

1) STRATOSPHERIC WAVE REFLECTION

In this section, we investigate the patterns of $v'T'$ and WAF to diagnose differences in large-scale wave forcing between AkR-, ArH-, and WCR-CAOs. We first focus on $v'T'$ at 100 hPa, as this variable is a proxy for vertically propagating waves into the stratosphere. Figures 7a, 7c, and 7e show the 100-hPa $v'T'$ for the AkR-, ArH-, and WCR-

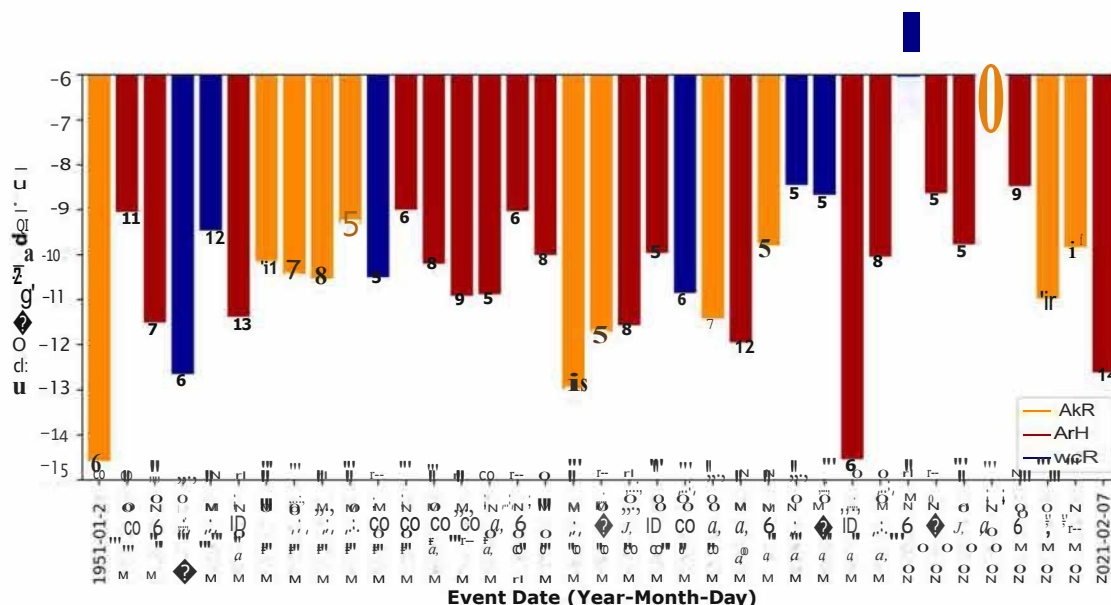


FIG. 6. Magnitude ($^{\circ}\text{C}$) of each of the 37 defined CAOs using the nondetrended TIM anomalies, categorized by weather regime: AkR (orange), ArH (red), and WCR (blue). Bold numbers below the bars represent the length (days) of each of the CAOs.

TABLE 1. Statistics for the mean and standard deviations of the magnitude and lengths of the CAOs by North American weather regime.

CAO regime type	Magnitude ($^{\circ}\text{C}$)	Magnitude standard deviation ($^{\circ}\text{C}$)	Length (days)	Length standard deviation (days)
AkR ($n = 12$)	-10.7	± 1.78	7.3	± 2.87
ArH ($n = 18$)	-10.5	± 1.52	8.1	± 2.76
WCR ($n = 7$)	-9.6	± 1.75	6.3	± 2.37

CAOs, respectively, averaged from days -3 to 0. For AkR- and ArH-CAOs, $\nu'T'$ is positive and significant over eastern Siberia. ArH-CAOs feature a slightly greater positive heat flux/vertical wave propagation signature over this region than the AkR-CAOs, with another region of significant positive $\nu'T'$ east of Greenland (Fig. 7c). Statistically significant positive $\nu'T'$ is actually present over Siberia during ArH-CAOs up to seven days before CAO onset (Fig. S3). Conversely, the strong positive $\nu'T'$ over the Siberian region is significant but only seen immediately prior to AkR-CAO onset (Fig. 7a and Fig. S4). Another notable feature is the negative 100-hPa $\nu'T'$ over northwest North America (Figs. 7a,c); this feature is much stronger for AkR-CAOs than ArH-CAOs. Together with the Siberian positive $\nu'T'$, this heat flux dipole is a hallmark signature for stratospheric wave reflection events over the PNA region (Matthias and Kretschmer 2020), indicating the presence of downward propagating/possibly reflected waves over North America. Finally, the WCR-CAOs do not feature a dipolar signature, with a lack of statistically significant signals over both Siberia and northwest North America (Fig. 7e).

Based on the zonal $\nu'T'$ dipole found in ArH- and AkR-CAOs, we modified the original RI from Matthias and Kretschmer (2020) by adjusting the boxes to focus on the regions of significant $\nu'T'$ over Siberia and Canada (black boxes in Figs. 7a,c,e). This index uses the standardized anomalies of the 100-hPa $\nu'T'$ over Canada subtracted from that over Siberia (see Fig. S5 for day -3 to 0 $\nu'T'$ anomalies only). The composites of the time-lagged modified RI are shown in Figs. 7b, 7d, and 7f for AkR-, ArH-, and WCR-CAOs, respectively. AkR-CAOs feature a rapid rise in the RI starting at day -8, peaking at a value of 2 one day prior to CAO onset (Fig. 7b). Analogous to the conclusions from the $\nu'T'$ composites (Fig. S3), the RI for the ArH-CAOs exceeds zero about a week prior to CAO onset, reaching a statistically significant value of 1.2 on day -2 (Fig. 7d). The ArH-CAO RI features a more gradual increase toward its peak, alongside smaller negative $\nu'T'$ in northwest North America (Fig. 7c) than in AkR-CAOs. This gradual increase in RI indicates more constant and prolonged vertical wave propagation from Siberia in ArH-CAOs. As such, ArH-CAOs have approximately 800 Km s^{-1} more cumulative total $\nu'T'$ at 100 hPa over Siberia than AkR-CAOs by day 0 (Fig. S6). This difference in cumulative $\nu'T'$ suggests a greater weakening of the SPY during ArH-CAOs versus AkR-CAOs, which is later verified via composites [section 3b(3)]. The WCR-CAOs feature no statistically significant signal prior to CAO onset due to the lack of significant $\nu'T'$ in the RI dipole region (Figs. 7e,f). As such, the stratosphere has a minimal

influence on WCR-CAOs. Since AkR- and ArH-CAOs appear to have more connections to the polar stratosphere and that the WCR-CAOs make up only $\sim 18\%$ of the total CAO sample size, we focus on the characteristics and formation of the dominant AkR- and ArH-CAOs for the remainder of this work.

The statistically significant peaks in the RI for AkR- and ArH-CAOs (Figs. 7b,d) emphasizes the question of whether stratospheric wave reflection plays a role in their development. Matthias and Kretschmer (2020) defined wave reflection events when the RI value exceeds 1.5 for 10 consecutive days. However, for our shorter-lived CAO events, the RI peaks immediately prior to CAO onset with notable asymmetry in the magnitudes of each pole of the index by regime type, especially for AkR-CAOs. The magnitude of positive $\nu'T'$ (i.e., upward wave propagation) from Siberia is slightly larger for ArH-CAOs than for AkR-CAOs. Also, the magnitude of negative $\nu'T'$ (i.e., downward propagating/possibly reflected waves) in northwest North America is much greater for AkR-CAOs than ArH-CAOs. These two observations are also shown in the anomalous $\nu'T'$ fields in Fig. S5. Since the RI measures the *difference* between the 100-hPa $\nu'T'$ standardized anomalies in these two areas, the index alone cannot determine if a reflective event is occurring. To clarify the high-latitude mechanism occurring in the days leading to Great Plains CAOs, Fig. 8 shows the lagged composites of the averaged 60° - 70°N linearly detrended GPH standardized anomalies (shaded contours) and full-field WAF (vertical and zonal components; vectors) for AkR- and ArH-CAOs. The AkR-CAOs feature persistent and significant anomalous stratospheric troughing over Eurasia through to day -4 (Figs. 8a,b). Simultaneously, there is anomalous stratospheric and tropospheric ridging over Alaska, alongside upward WAF propagating into the stratosphere over eastern Eurasia and Siberia (100° - 160°E). Figure 8b indicates downward propagating WAF from the stratosphere over Alaska and into northwest Canada, with upward WAF from the surface to the midtroposphere. Finally, Fig. 8c shows amplification of the anomalous ridging over Alaska and the wave fluxes seen in the previous lags, with notable stratospheric wave reflection occurring between days -3 and 0. The wave reflection is denoted by statistically significant WAF propagating from the troposphere to the stratosphere over Siberia, alongside statistically significant downward WAF from the stratosphere to the troposphere over Canada. This panel corresponds to the peak in the AkR-CAO RI (Fig. 7b). The evident midtropospheric WAF convergence results in troughing over Canada, decelerating the westerly flow, and amplifying the midtropospheric wave pattern (Perlitz and Hamik 2003).

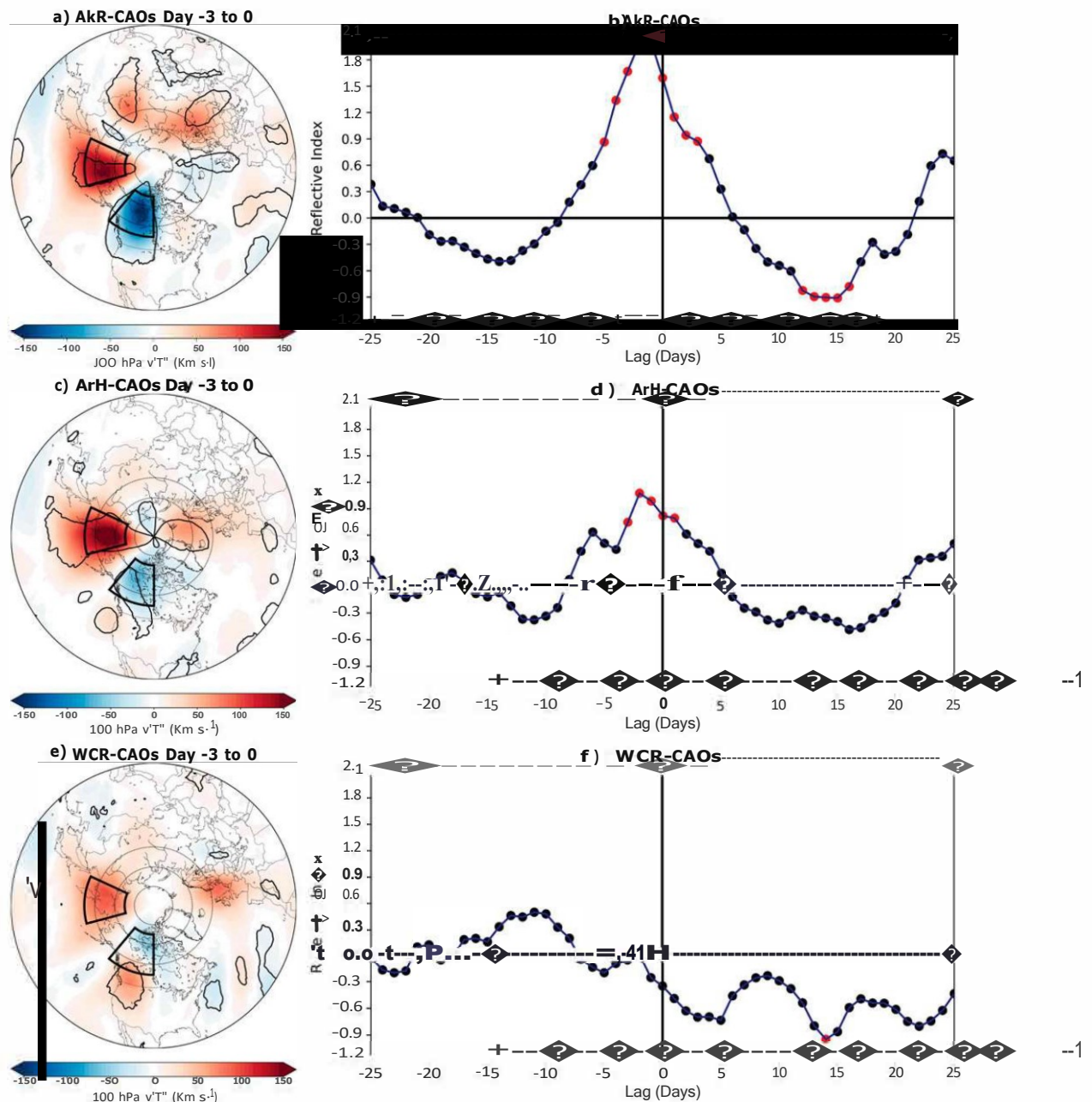


FIG. 7. Composites of (a),(c),(e) 100-hPa $v'T''$ (Km s⁻¹, shaded) from days -3 to 0 and (b),(d),(f) RI for AkR-, ArH-, and WCR-CAOs, respectively. The RI is defined as in Matthias and Kretschmer (2020), but between 55°–75°N and 145°–185°E and 220°–260°E for Siberia and Canada, respectively [thick black lines in (a), (c), and (e)]. Black contours in (a), (c), and (e) and red dots in (b), (d), and (f) show statistical significance ($p < 0.05$) via 5000-iteration bootstrapping. The two southerly dashed lines are latitudes 60° and 70°N, as used in Fig. 8. Negative and positive lags indicate days before and after CAO onset day (day 0), respectively.

The ArH-CAOs show a different overall evolution of wave activity. Significant anomalous tropospheric troughing develops over Eurasia between days -11 and -4, along with anomalous stratospheric troughing between days -11 and -8, and upward WAF from the upper troposphere into the stratosphere over Siberia (Figs. 8d,e). Significant anomalous tropospheric ridging already exists near 300°E from days -11 to -8 (Fig. 8d), earlier than the appearance

of the anomalous ridging for AkR-CAOs (Figs. 8a-c). This anomalous ridging intensifies toward CAO onset and retrogrades westward (Fig. 8f). Importantly, the canonical wave reflection pattern is not seen during ArH-CAOs. Therefore, while the RI indicates a statistically significant peak immediately before both ArH- and AkR-CAOs (Fig. 7), stratospheric wave reflection only occurs during AkR-CAOs (Fig. 8c).

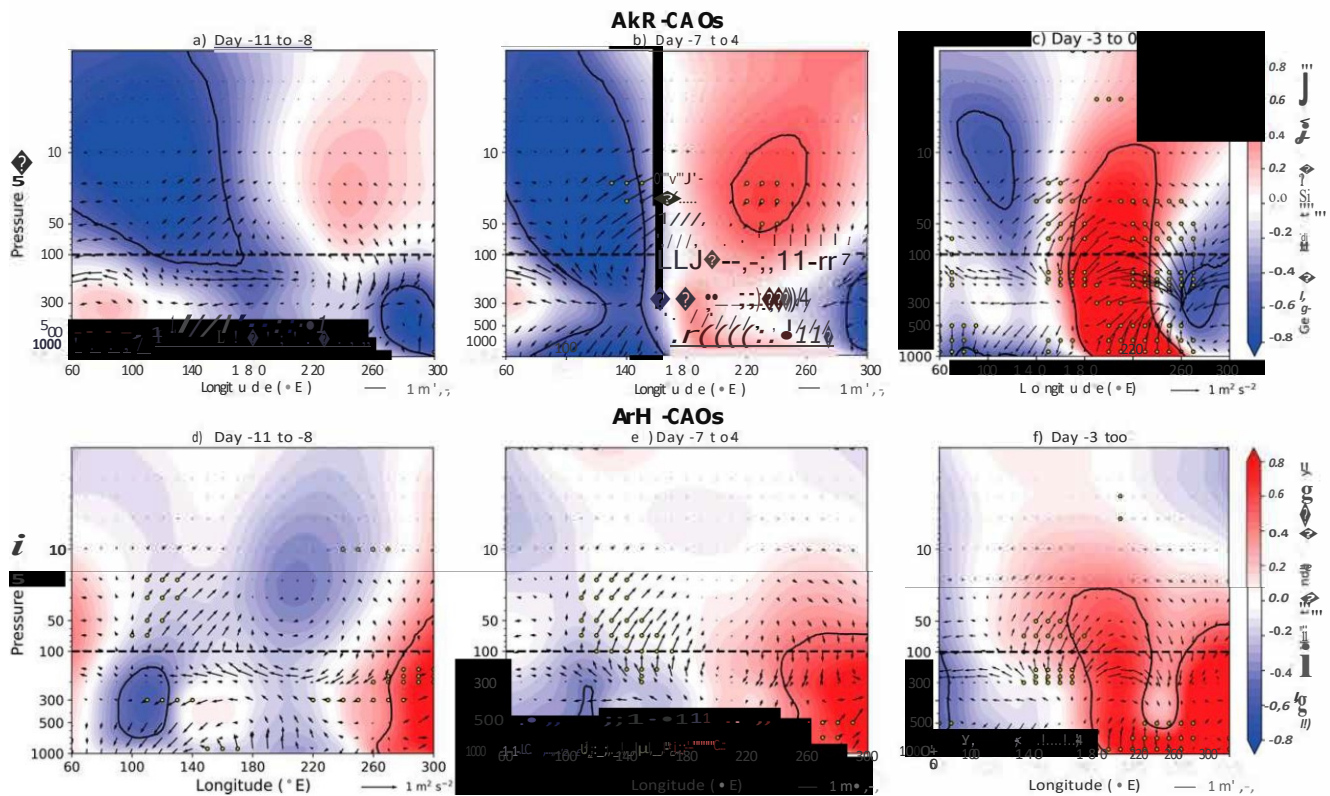


FIG. 8. Lag composites of the linearly detrended standardized anomalies of GPH (shaded) and full-field WAF (vectors) at 4-day intervals from days -11 to -8 through days -3 to 0 for (a)–(c) AkR-CAOs and (d)–(f) ArH-CAOs, averaged between 60° and 70°N . Yellow dots represent statistical significance ($p < 0.05$) of the vertical component of the WAF via a 5000-iteration bootstrapping without replacement. Black contours represent statistical significance ($p < 0.05$) of the GPH anomalies using the same method. Negative lags indicate days prior to CAO onset day (day 0), and positive lags indicate days after CAO onset. Scaling of the WAF vectors was achieved by dividing the vertical WAF component by a scale factor of 0.1 above 200 hPa and by a scale factor of 0.3 below and including 200 hPa, and dividing the zonal WAF component by 10 above 200 hPa and by 50 below and including 200 hPa.

Horizontal wave propagation differences between the AkR- and ArH-CAOs are also apparent across the Pacific domain. Figure 9 shows the lag-composited 250-hPa streamfunction anomalies and only the statistically significantly large ($p < 0.05$) standardized WAF (meridional and zonal) anomalies in this region prior to and including CAO onset. During AkR-CAOs (Figs. 9a,c,e), a Pacific wave train develops from the subtropical Pacific to North America and intensifies toward CAO onset. This wave train is not seen for ArH-CAOs (Figs. 9b,d,f). Instead, there are significant streamfunction and WAF anomalies in the North Atlantic associated with ArH-CAOs (Figs. 9b,d,f). These findings suggest that precursor patterns leading to AkR-CAOs in Figs. 8 and 9 may initiate through tropical remote forcing. Indeed, previous studies have linked certain phases of the Madden-Julian oscillation (MJO), and El Niño-Southern Oscillation (ENSO), to wave breaking and high-latitude blocking in the Gulf of Alaska, setting up an AkR regime-like pattern (e.g., Renwick and Wallace 1996; Henderson and Maloney 2018). By contrast, ArH-CAOs feature pan-Arctic connections, likely tied to the weakened SPV and its downward influence on the NAM (e.g., Baldwin and Dunkerton 2001; Hitchcock and Simpson 2014).

2) CONDITIONS FOR WAVE REFLECTION DURING AKR-CAOs

In the previous section, we found that AkR-CAOs featured stratospheric wave reflection immediately prior to CAO onset (Figs. 8b,c). We now investigate the specific conditions that constitute wave reflection, as outlined in previous studies (e.g., Perlwitz and Harnik 2003, 2004; Matthias and Kretschmer 2020). Figures 10a and 10b show the vertical profile of area-averaged zonal wind between 70° – 80°N and 200° – 250°E (i.e., the area of downward reflection in Fig. 8c), and Figs. 10c and 10d show the meridional profile of the zonal wind averaged over the same longitudes at 30 hPa, analogous to Matthias and Kretschmer (2020). Here we composite days -7 to -4 and days -3 to 0 to represent the two panels in Fig. 8 where the reflection mechanism appears. Figures 10a and 10b show an anomalously strong zonal wind between 200° and 250°E throughout the stratosphere, and negative vertical wind shear above 10 hPa approaching CAO onset (solid red line). The zonal mean zonal wind profile also demonstrates slightly stronger zonal winds in the stratosphere (Figs. 10a,b, solid blue line). Increased curvature in the regional vertical profile is consistent in the mid-to-upper stratosphere leading up to CAO onset

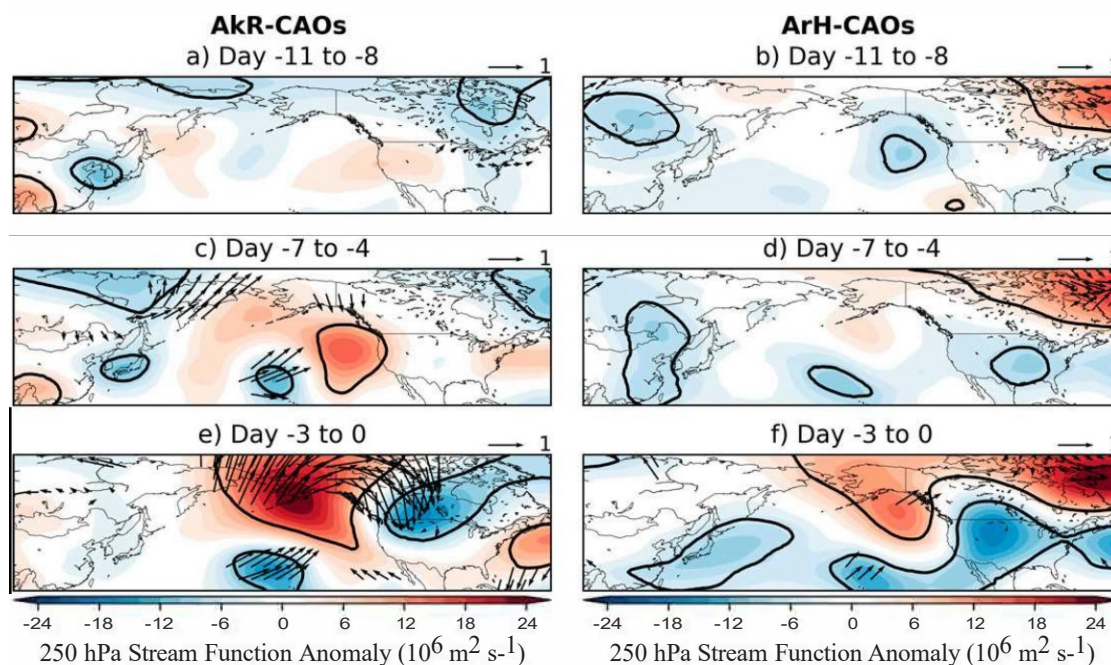


FIG. 9. Lag composites of the 250-hPa streamfunction anomalies ($10^6 \text{ m}^2 \text{ s}^{-1}$; shaded) and statistically significant ($p < 0.05$) standardized WAF anomalies (vectors) across the Pacific basin for 4-day intervals from days -11 to -8 through days -3 to 0 for (a),(c),(e) AkR-CAOs and (b),(d),(f) ArH-CAOs. Black contours represent statistical significance ($p < 0.05$) of the streamfunction anomalies via a bootstrapping approach of 5000 iterations without replacement. Statistical significance ($p < 0.05$) of WAF is computed using the magnitude of the WAF anomalies and 5000-iteration one-tailed bootstrapping without replacement. Negative lags indicate days prior to CAO onset day (day 0), and positive lags indicate days after CAO onset.

(Figs. 10a,b; solid red line and gray dots), with the zonal mean zonal wind vertical profile also showing increased curvature at the same lags (Figs. 10a,b; solid blue line and gray dots). The meridional profiles of the zonal wind (Figs. 10c,d) demonstrate increased curvature (gray dots) of the regional (solid red line) and zonal mean (solid blue line) zonal wind in both the midlatitudes and high latitudes approaching CAO onset, indicating a meridional waveguide between approximately 45° – 80°N . The regionally averaged meridional profile (Fig. 10c, solid red line) is most striking, with a well observed anomalous peak in zonal winds in the high-latitude stratosphere with increased curvature, a condition required for downward coupling from wave reflection (Matthias and Kretschmer 2020). Hence, this analysis suggests that the AkR-CAOs satisfy the wave reflection conditions.

3) RELATIONS TO SPV VARIABILITY

Given the strong $\mathbf{v}' \cdot \mathbf{T}'$ and WAF signature over Siberia for ArH-CAOs (Figs. 7c,d and 8e,f), we next investigate the role that SPV variability may have on the development of ArH-CAOs. SPV variability is of importance since strong or weak SPV states have implications for S2S predictability—e.g., the downward propagation of NAM anomalies (e.g., Baldwin and Dunkerton 2001; Tripathi et al. 2015; White et al. 2019). Figure 11 shows pressure-time lag composite cross sections of the standardized anomalies of the zonal wind (Figs. 11a,c) and linearly detrended GPH standardized anomalies (Figs. 11b,d) for AkR- and ArH-CAOs, both area averaged over the polar

cap (60° – 90°N). Here, we perform lag compositing from days -40 to +40 (day 0 is the CAO onset date) since the polar stratosphere has significantly longer time scales for its evolution. Figure 11a illustrates that AkR-CAOs are preceded by statistically significantly ($p < 0.05$) stronger than average polar cap zonal winds 1–3 weeks before CAO onset throughout much of the troposphere and stratosphere up to 10 hPa. These anomalously strong winds are accompanied by anomalously negative GPH anomalies throughout the stratosphere and upper troposphere (Fig. 11b). Interestingly, there is a very sharp transition to weaker than average zonal winds and more positive GPH anomalies just before day 0, supporting the result that AkR-CAOs undergo a rapid transition after wave reflection, associated with stronger SPV conditions (Figs. 8c, 11b, and Kretschmer et al. 2018a).

The patterns of zonal wind and GPH anomalies seen in AkR-CAOs are in stark contrast to those for ArH-CAOs (Figs. 11c,d). The lag composites of both the zonal wind anomalies and GPH anomalies for ArH-CAOs present a classic "dripping paint" signature, i.e., downward propagation of negative zonal wind anomalies and positive GPH anomalies from the stratosphere into the troposphere (Baldwin and Dunkerton 2001). Figure 11c shows significantly weaker than normal zonal winds in the upper stratosphere between days -30 and -40. This signal propagates downward into the mid-to-lower stratosphere between days -30 and -10, before reaching the surface near CAO onset. Downward propagation appears in the GPH anomalies for the ArH-CAOs as

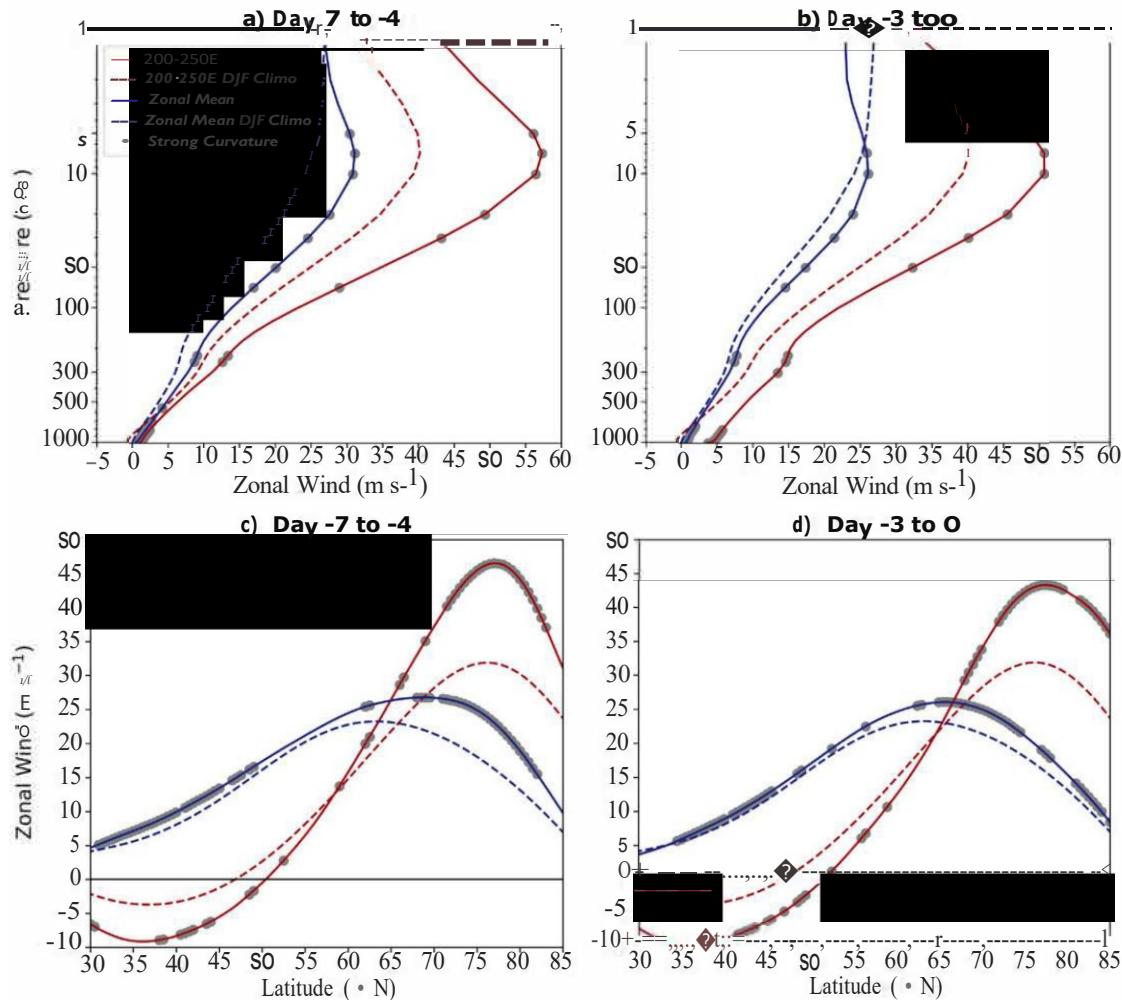


FIG. 10. Lagged composites of the (a), (b) vertical profile averaged between 70° and 80°N, and meridional profile at 30 hPa, (c), (d) of zonal wind (m s^{-1}) for AkR-CAOs averaged between 200° and 250°E (red solid line) and in the zonal mean (blue solid line), at 4-day intervals from days -7 to -4 and from days -3 to 0, respectively, where day 0 is CAO onset. Red and blue dashed lines represent DJF 1981-2010 climatology of the zonal wind between 200° and 250°E and in the zonal mean, respectively. Gray dots indicate strong curvature (see section 2d). Negative lags indicate days prior to CAO onset day (day 0), and positive lags indicate days after CAO onset.

well (Fig. 11d), with statistically significant downward propagation of positive GPH anomalies to the surface from around 50 hPa in the two weeks preceding these events. Although not statistically significant, the downward propagation of these GPH anomalies can be seen higher in the stratosphere on longer time scales, consistent with negative zonal wind anomalies (Fig. 11c). A possible explanation is that while a weak SPV is a favorable precursor to Great Plains CAOs, a *major* SSW may not be necessary. In support of this explanation, only seven of the ArH-CAOs were preceded by a major SSW in the period of up to 6 weeks prior to CAO onset [see Butler et al. (2017) and Afargan-Gerstman and Domeisen (2020) for a list of dates], including the *major* SSW in January 2021 (Lee 2021). No major SSWs preceded AkR- and WCR-CAOs.

The results of Fig. 11 clearly demonstrate that the state of the SPV influences the formation of CAOs in the Great Plains, but its exact dynamical influence depends on the regime type at CAO onset. Such an observation now begs the

question of whether the influence of the SPV on CAO formation can be diagnosed in advance, to determine whether a CAO will occur with a given onset day regime. To assess this prospect, Fig. 12 shows North American weather regime transitions between day -20 and day +10 for both AkR- and ArH-CAOs. First, AkR-CAOs mainly show a transition to the AkR regime on synoptic time scales (i.e., a few days), from persistent WCR, PT, and/or ArL states (Fig. 12a). PT and ArL regimes are associated with anomalously negative GPH anomalies in the Arctic stratosphere, indicating a stronger SPV (Lee et al. 2019). Therefore, AkR-CAOs preferentially originate from regimes associated with stronger SPV conditions, aligning with the results in Fig. 11a. The WCR regime has a closely located anomalous ridging node to the AkR, also making it a preferential precursor regime (Figs. 4a,d). Thus, persistent periods of WCR, PT, and/or ArL regimes could increase the probability of an AkR-CAO in the U.S. Great Plains several days to weeks later.

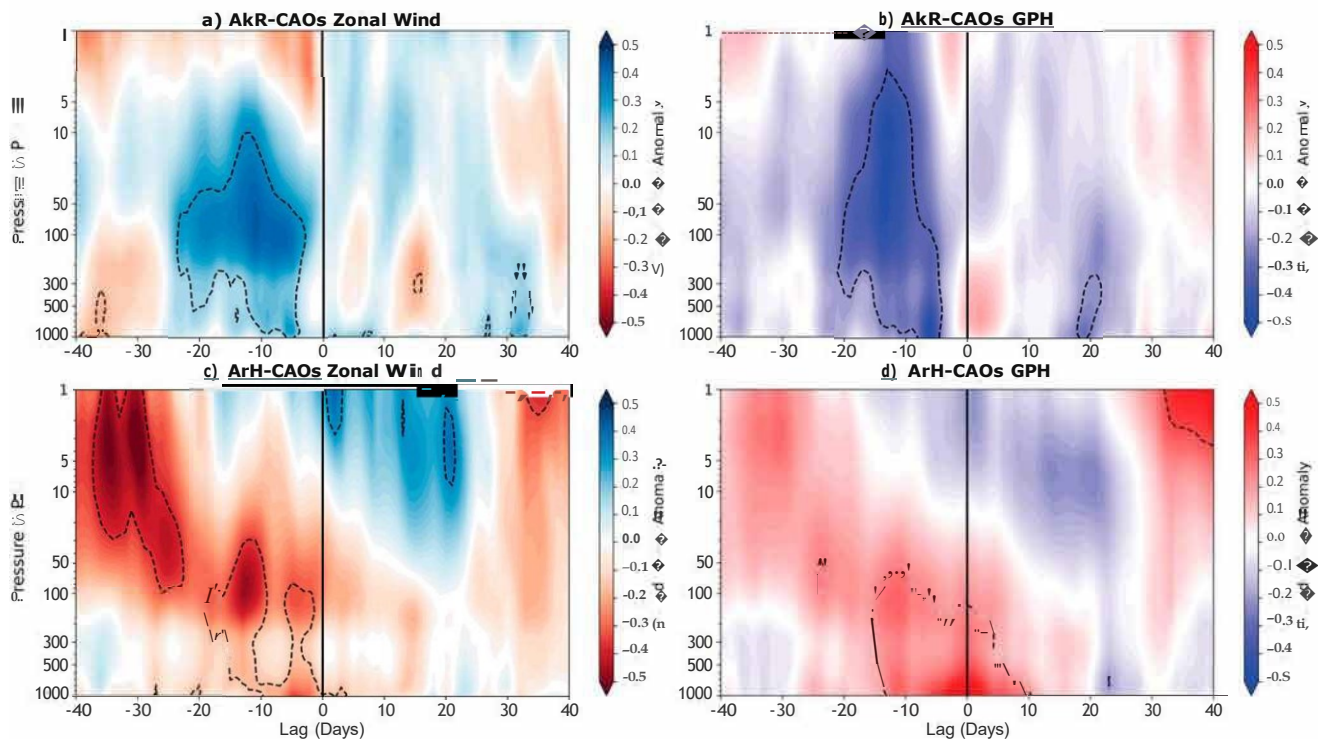


FIG. 11. C composite pressure-time polar cap (60° – 90° N) averaged cross sections of standardized and nondetrended zonal wind anomalies for (a) AkR- and (c) ArH-CAOs, and of linearly detrended GPH standardized anomalies for (b) AkR- and (d) ArH-CAOs between days -40 and +40. Dashed lines represent statistical significance at the 95th percentile via bootstrapping of 5000 iterations. Negative lags indicate days prior to CAO onset day (day 0), and positive lags indicate days after CAO onset.

The persistence of the ArH regime prior to ArH-CAO onset is visualized in Fig. 12b. In most cases, the ArH regime is the dominant regime for approximately a week or longer before CAO onset. ArH regime days are associated with anomalously positive GPH anomalies in the Arctic stratosphere (Lee et al. 2019), yielding an anomalously weak SPV. The persistent signal of the ArH regime before ArH-CAO onset relates to the results of Figs. 11c and 11d. The downward propagation of negative zonal wind anomalies and

positive GPH anomalies eventually yields a tropospheric ridging signal in the Arctic troposphere within the 20 days preceding CAO onset. However, is it possible that such a scenario could happen without a CAO? To answer this, we investigated events with 5 or more consecutive ArH days that were preceded by 1 4-day averaged standardized zonal wind anomalies poleward of 60° N of less than -0.3 at 100 hPa, and/or less than -0.4 at 10 hPa 25–35 days prior to the last ArH day (see Fig. 11c). Out of a total of $n = 43$ such events, 8 of the

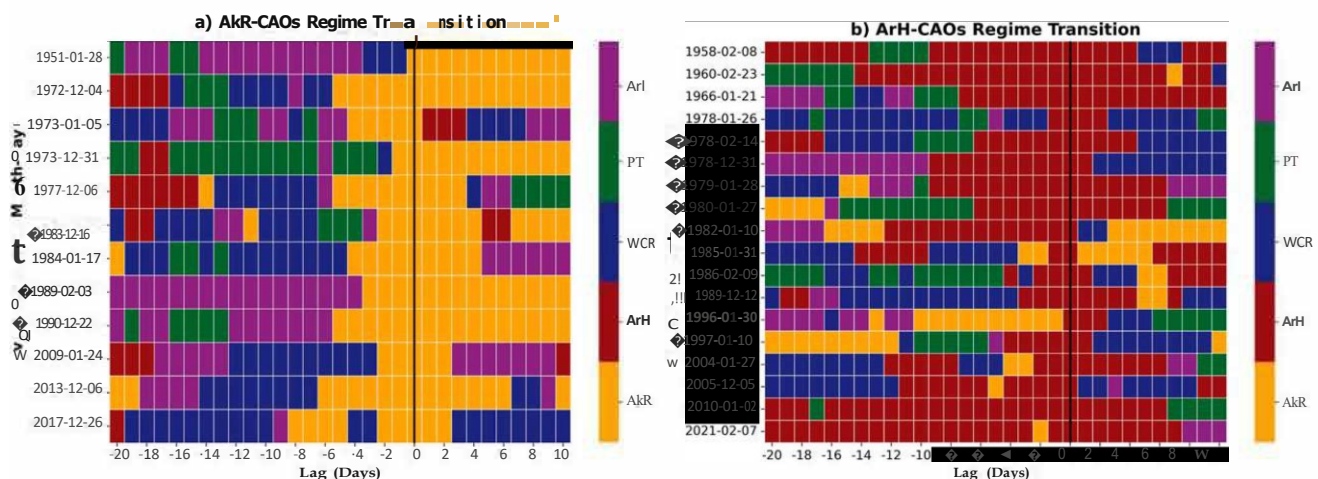


FIG. 12. North American weather regime transitions for (a) AkR-CAOs and (b) ArH-CAOs in the U.S. Great Plains, between days -20 and +10. Negative lags indicate days prior to CAO onset day (day 0), and positive lags indicate days after CAO onset.

ArH-CAOs fall into this category. The other 35 identified events do not meet our CAO criteria but indeed feature TIM anomaly patterns analogous to Fig. 1 with cold anomalies propagating from the northwest, but with weaker magnitudes of $\sim 5^{\circ}\text{C}$ in the Great Plains. Thus, a weak SPY followed by persistent ArH regimes increases the risk of cold periods for the Great Plains.

4. Discussion

Wintertime CAOs in the U.S. Great Plains are high-impact events, though few studies currently exist about their formation and potential S2S predictability. This work fills part of this gap. Consistent with previous studies, we show that CAOs in the Great Plains are associated with two dominant weather regimes: anomalous ridging over Alaska or anomalous ridging over Greenland and northern Canada (Figs. Sa,b). Subsequently, the dynamics leading up to these two CAO types are unique and act on different time scales. The differing dynamical characteristics found between different regime-type CAOs demonstrate that using a regime-based approach to classify the CAOs is an efficient, computationally inexpensive, and powerful tool in distinguishing between key differences in dynamical forcings for extreme events.

The AkR-CAOs feature a rapid development of anomalous ridging over Alaska with a wave train across the North Pacific and anomalous horizontal WAF from the central Pacific, associated with a short time scale wave pulse from Siberia leading to stratospheric wave reflection (Figs. 8a-c and 9). Thus, Great Plains AkR-CAOs are sensitive to changes in upstream GPH patterns across the North Pacific, which themselves have strong teleconnections to MJO activity (e.g., Renwick and Wallace 1996; Henderson and Maloney 2018; Green and Furtado 2019). The rapid development of the AkR-CAOs (i.e., in a few days) aligns with previous studies which note that stratospheric wave reflection generally occurs in association with rapidly developed wave activity into the stratosphere (Harnik 2009; Kodera et al. 2016; Kretschmer et al. 2018a; Matthias and Kretschmer 2020).

In contrast, ArH-CAOs feature large-scale weak SPY conditions propagating downward over subseasonal time scales (Figs. 11c,d), with both weak upper and lower SPY conditions favored. These results complement past studies on the downward influence of the stratosphere on the troposphere via projecting onto the negative phase of the NAM, which is associated with the southward migration of the polar jet stream and an enhanced risk of Northern Hemisphere CAOs (e.g., Baldwin and Dunkerton 2001; Limpasuvan et al. 2004; Hitchcock and Simpson 2014; White et al. 2019; Maycock et al. 2020). While previous literature suggests that SPY variability is not a robust precursor *on its own* for U.S. CAOs (e.g., Cellitti et al. 2006; Smith and Sheridan 2018b), our study suggests that the precursor potential of SPY variability to Great Plains CAOs relates to the North American winter weather regime in place in the troposphere. Weak SPY conditions and persistent ArH regimes prior to CAO onset relate to ArH-CAOs, while strong SPV conditions and persistent stronger weather regimes relate to the AkR-CAOs (Figs. 11a,c, and 12).

Moreover, ArH-CAOs feature a longer time scale downward propagation of negative zonal wind anomalies up to around 5 weeks before CAO onset. However, the stronger SPY before AkR-CAOs is seen up to 3 weeks before CAO onset. Thus, the ArH-CAOs feature a longer precursor time scale, potentially resulting in long-lead S2S predictability.

Interestingly, our modified RI identifies significant peaks for AkR- and ArH-CAOs immediately before onset, but with the AkR-CAOs showing evolution over synoptic time scales (e.g., 5 days or less) (Figs. 7b,d). However, ArH-CAOs did not demonstrate the reflective trait but instead had intense Siberian positive 100-hPa $v'T'$ anomalies which lacked the focused dipole in AkR-CAOs, in association with deeper negative anomalies over Canada. The disparity between RI values and the lack of wave reflection for ArH-CAOs shows that one should examine spatial maps of the 100-hPa $v'T'$ alongside the RI. Using both tools shows that the SPY can absorb the intense Siberian wave activity in ArH-CAOs, aiding the weak vortex regime, but the stronger SPV preceding AkR-CAOs more readily facilitates wave reflection (Figs. 8 and 11).

The relationship of the CAOs with SPY variability also has climate change implications. That is, the frequency and intensity of these events in a future warmer world could be sensitive to climate change-induced effects on the strength and position of the SPY, as well as on the tropospheric circulation patterns. There has been some uncertainty surrounding the exact impact of anthropogenic climate change on the SPV, with future climate simulations showing no consensus on *major* SSW frequency, and even a decrease in weak SPY downward coupling signals (Ayarzaguena et al. 2020; Hamouda et al. 2021). However, recent studies have shown that the frequency of weak SPY states has increased in the Arctic amplification era, especially the occurrence of "SPY stretching" events (Kretschmer et al. 2018b; Cohen et al. 2021). These "SPY stretching" events are often associated with Alaskan ridging and a dipole in WAF analogous to that of the 100-hPa $v'T'$ in Fig. 7a. Therefore, an increase in these events could increase AkR-CAO risk in a warming climate. Future work will investigate the frequency changes of AkR and ArH regimes under future climate change, their relationship to the SPY, and how these trends affect CAOs across North America, particularly the Great Plains.

Finally, although the formation of the reflective mechanism associated with AkR-CAOs is fast, there may be potential for anomalous Alaskan ridging to be linked with the tropics. The MJO and ENSO have already been linked with the development of North Pacific blocking, with previous studies showing 1) an increase in North Pacific wintertime blocking during El Niño when followed by MJO phase 7, and 2) increased sensitivity of blocking frequency in the Bering Strait to the ENSO cycle (Renwick and Wallace 1996; Henderson and Maloney 2018). Furthermore, Green and Furtado (2019) showed that MJO phases 7 and 8, alongside a strong SPV, aids the development of anomalous tropospheric ridging over Alaska. Conversely, a weak SPY alongside several MJO phases correspond to anomalous Arctic ridging. Further work should assess MJO phase dependence and ENSO modulation for Great Plains CAOs, and their relationships with the SPY and high-latitude

blocking. These lines of research are important for assessing whether these modes of variability could also act as key S2S predictors for the fast time scale processes involved in AkR-CAOs and other extreme weather events.

5. Conclusions

In this study, 37 U.S. Great Plains CAOs were defined and analyzed between 1950 and 2021. The main conclusions were as follows:

- The CAOs had an average duration of 7.5 days and an average TIM anomaly of -10.4°C across the domain. These events occurred approximately once every two winter seasons on average.
- The dominant onset day regimes for these CAOs were the Alaskan ridge (AkR; Figs. 4a and Sa) and the Arctic high (ArH; Figs. 4b and Sb). ArH-CAOs are slightly longer and AkR-CAOs slightly colder, although these differences are not significant (Table 1).
- AkR-CAOs involve stratospheric wave reflection, consisting of upward wave activity from Siberia alongside downward wave activity over Canada, and a stronger than average SPV preceding the event (Figs. 8c, 11a,b).
- CAOs that start with an ArH regime feature a longer time-scale buildup of positive heat flux from Siberia (Fig. 7c and Fig. S3), accompanied by a weak upper and lower SPV, featuring downward propagation of zonal wind and GPH anomalies (Figs. 11c,d).

A caveat to this work is that using domain-averaged TIM anomalies to define CAOs may mean that extremely cold conditions do not always occur throughout the entire region. Future work should consider splitting the study area into the southern and northern Great Plains and studying each separately. Another way to define these events may involve implementing a spatial extent threshold to identify more regional Great Plains CAOs, as in Smith and Sheridan (2020). The differences in the time scale and evolution between these two dominant types of CAOs (i.e., AkR- and ArH-CAOs) invites future research. One particular caveat that should be addressed is the further investigation of the fraction of times in regimes and weak/strong SPV conditions in which CAOs do not occur, to further enable application to S2S prediction. Additionally, in the light of recent work by Lillo et al. (2021) and Biernat et al. (2021), future work should investigate any potential link with tropopause polar vortices migrating within the large-scale flow.

Another avenue of future work is identifying potential S2S predictability windows of opportunity for Great Plains CAOs involving tropical to extratropical interactions of the MJO and ENSO. The rapid Alaskan ridge development could be linked to Rossby wave breaking from convection in the tropics, potentially modulated by ENSO (Renwick and Wallace 1996; Moore et al. 2010; Seo et al. 2016; Seo and Lee 2017; Henderson and Maloney 2018). The investigation of how predictable these CAOs are in S2S model hindcasts is also underway. Additionally, studying the formation, evolution, and dynamics of Greenland and Alaskan blocking may

provide further sources of S2S predictability for Great Plains CAOs. Understanding these high-latitude blocking areas is important as they could portend CAOs in North America, especially since weather models inherently struggle representing their frequency and longevity (Quinting and Vitart 2019). Investigating these concepts carries great potential for improving the predictability of these CAOs on the S2S time scale, and for enhancing impact mitigation.

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Data availability statement. ERAS reanalysis data can be found in the Copernicus Climate Data Store at <https://cds.climate.copernicus.eu/#!/home>.

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